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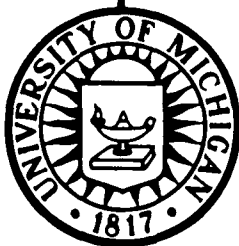
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VESIAC State-of-the-Art Report

SEISMOLOGICAL EVIDENCE PERTAINING TO THE STRUCTURE OF THE EARTH'S UPPER MANTLE

OTTO NUTTLI

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ACOUSTICS AND SEISMICS LABORATORY

Institute of Science and Technology

THE UNIVERSITY OF MICHIGAN

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Contract SD-78

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OTTO NUTTLI

Associate Professor
St. Louis University

January 1963

Acoustics and Seismics Laboratory

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THE UNIVERSITY OF MICHIGAN

Ann Arbor, Michigan

NOTICES

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PREFACE

VESIAC, the VELA Seismic Information Analysis Center, is an information collection, analysis, and dissemination facility established at the Institute of Science and Technology of The University of Michigan. The contract is sponsored by the Advanced Research Projects Agency under the Office of the Secretary of Defense.

The purpose of VESIAC is to analyze the research information related to the VELA UNIFORM Program of Project VELA and to function as a central facility for this information. The facility will serve all authorized recipients of VELA UNIFORM research information by issuing subject bibliographies with abstracts, annotated bibliographies, and special reports as required. In addition, VESIAC will periodically summarize the progress of the research being conducted.

VESIAC is under the technical direction of the Acoustics and Seismics Laboratory of the Institute. In its operation VESIAC draws upon members of this laboratory and other members of the Institute and University.

ACKNOWLEDGEMENTS

This report was written at The University of Michigan during the summer of 1962, while the writer was on sabbatical leave from Saint Louis University. He wishes to thank all the members of the Acoustics and Seismics Laboratory and of the VESIAC staff of the Institute of Science and Technology of The University of Michigan for making his visit such a pleasant one. In particular, he wishes to express his gratitude to Dr. James T. Wilson, acting director of the Institute of Science and Technology, Mr. Thomas Caless, research associate and principal investigator of VESIAC, and Mr. R. C. Haven, research associate. He wishes to acknowledge the contribution of Dr. Carl Kisslinger of Saint Louis University and Mr. Henry Pollack of The University of Michigan, both of whom read the manuscript and made valuable comments. He also wishes to thank Dr. Carl Romney of AFTAC and Dr. Eugene Herrin of Southern Methodist University for sending prepublication copies of their manuscripts and for permission to use them in this report.

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SEISMOLOGICAL EVIDENCE PERTAINING TO THE STRUCTURE OF THE EARTH'S UPPER MANTLE

ABSTRACT

This state-of-the-art report concerns the variation of the velocity of P and S waves with depth in the upper mantle of the earth, and the effect of this variation upon the recorded earth motion at distances less than 3000 km from the source. Body-wave and surface-wave study results of many authorities are examined, and the following conclusions presented: (1) there are significant regional variations of velocity in the upper mantle, (2) there is no pronounced shadow zone for P waves resulting from a "low-velocity channel" in the upper mantle, except in tectonically active regions such as California-Nevada, Japan, Andes, etc., and (3) there is a significant, world-wide "low-velocity channel" for S waves. A tabulated summary of upper mantle structure according to geographical region is presented in Table I. The report concludes with a summary, recommended problems for future research, and a statement of instrumentation requirements for further study of the upper mantle.

1

STATEMENT OF THE PROBLEM

One of the oldest and most fundamental seismological problems is the determination of the velocities of elastic body waves, P and S, as a function of depth within the earth. Great advances were made in the solution of this problem about twenty-five years ago when Gutenberg and Richter [1] and Jeffreys [2] computed values for these velocities on the basis of the observed travel times of earthquake-generated waves. It is significant that the two sets of values of velocity for the mantle (the region of the earth between depths of approximately 35 and 2900 km) were within 10% of one another at any given depth, and were nearly identical at depths greater than 500 km.

Further study has shown that these differences, although quantitatively small, are quite significant. Because of its importance, the problem (P and S velocities as a function of depth) continues to be a subject of research for many of the world's most competent geophysicists. The Executive Committee of the International Union of Geodesy and Geophysics, at its 1960 triennial meeting in Helsinki, Finland, recognized the importance of such study and gave the Union's sponsorship to a research plan of international participation entitled "The Upper Mantle and Its Influence on the Development of the Earth's Crust."

The upper mantle is bounded above by the Mohorovicic discontinuity, which separates it from the crust. The lower limit is not so well defined, but for the purpose of this report it is considered to be at a depth of approximately 500 km. Some of the reasons for this choice become apparent in Sections 2 and 3. However, two may be mentioned at this time. First, the velocity distributions in the mantle (as determined by Jeffreys and by Gutenberg and Richter) are practically identical below 500 km; therefore the upper mantle by definition will include all of that portion in which there is some disagreement concerning the velocity-depth relation. Second, the study of the group velocity of Rayleigh waves by Dorman, Ewing, and Oliver [3] indicates that there are regional, or lateral, changes in velocity down to a depth of about 500 km, but below this level the velocity is only a function of the depth coordinate.

It is important to have a precise knowledge of the variation of the P- and S-wave velocities, both lateral and vertical, in the upper mantle. This region of the earth possibly holds clues to the basic and difficult problems of orogenesis, which may be inferred from the complete title of the "Upper-Mantle Project" (as it is frequently called) of the International Union of Geodesy and Geophysics. Also, the structure of the upper mantle influences both the travel times and the amplitudes of P and S waves at distances from a few hundred to about 3000 km, and thus is highly significant in the problem of detecting and identifying underground nuclear explosions by seismological techniques. Accurate information of P- and S-wave velocity variation with depth will contribute heavily to such areas of research.

For practical reasons, the scope of this report will be limited to seismological evidence about the structure of the upper mantle. Unfortunately, the problem itself is not limited in this manner; vertical and lateral variation of the elastic body-wave velocities is intimately related to variations of temperature, chemical composition, density, distribution of radioactive elements, electrical and thermal conductivity, etc. This report, however, is concerned primarily with the information provided by seismology in determining the velocity variation in the upper mantle, and secondarily with the application of this knowledge to the interpretation of seismograms.

Since another VESIAC state-of-the-art report will be concerned with the structure of the earth's crust, and since the determination of the P_n and S_n velocities is intimately related to the crustal problem, this upper-mantle report will not consider the results of those seismological investigations containing P_n and S_n velocity determinations and devoted primarily to crustal structure studies. For practical purposes, this means that refraction studies carried out to distances no greater than 400 or 500 km will be excluded, as will surface-wave studies involving periods no greater than about 50 to 75 seconds.

Sections 2 and 3 of this report are concerned with body-wave and surface-wave studies that relate to the upper mantle. These sections attempt to present the results of the many inves-

tigators of the velocity-depth relationship problem, without reflecting the opinions of the present writer, although the effort will doubtless be only partially successful. Section 4 presents an analysis of the results of the research, and includes suggestions concerning problems which remain to be solved.

2

BODY-WAVE STUDIES

2.1. METHOD OF ANALYSIS

Travel time, amplitude, and period are body-wave properties most frequently used to determine the variation of velocity with depth within the earth. Before discussing observational results, it may be desirable to indicate briefly how these properties can be used for this purpose. A model of the earth is assumed for which the velocity of propagation of a body wave, v , is a function only of distance, r , from the center of the model. The wave may be either a P or an S wave.

2.1.1. VELOCITY-DEPTH VARIATION EFFECT ON TRAVEL TIMES AND AMPLITUDES. Both the variation of velocity with depth and the focal depth will affect the character of the travel-time curves and the amplitudes of the body waves. The following subsections present a qualitative discussion of such effects for those cases which appear to be most pertinent to a study of the structure of the earth's upper mantle.

2.1.1.1. "Ordinary" Behavior of Velocity. For most regions within the earth, velocity increases slowly and continuously as a function of depth. This behavior of velocity is called "ordinary." It is well known [4] that the travel-time curve resulting from such a velocity variation is concave downward, with the curvature decreasing as the distance increases (see Figure 1). The travel time of the body wave is represented by t , and Δ refers to the epicentral distance.



FIGURE 1. TRAVEL-TIME CURVE
FOR AN ORDINARY INCREASE OF
VELOCITY WITH DEPTH

The amplitude of the wave as it arrives near the surface of the earth¹ will be a slowly-decreasing function of Δ .

2.1.1.2. Rapid Increase in Velocity. Bullen [5] has considered the case in which the velocity increases rapidly with increasing depth beyond a certain depth, while above and below this level the behavior is "ordinary"; he has derived expressions for the travel times and amplitudes. As may be seen in Figure 2, this abnormal velocity variation results in a triplication of the travel-time curve. If the velocity increase is discontinuous, neither of the cusps at A and B will be associated with large amplitudes. If the velocity increase is continuous and the velocity gradient discontinuous, then the cusp at B will correspond to large amplitudes. If both the velocity and velocity gradient are continuous, then the cusps A and B will both be associated with large amplitudes.

2.1.1.3. Discontinuous Decrease in Velocity. Assume that the velocity decreases discontinuously in proceeding from above to below a certain depth, and then behaves ordinarily as the depth further increases. The travel-time relations derived by Bullen [5] are noteworthy

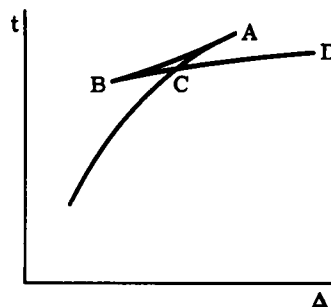


FIGURE 2. TRAVEL-TIME CURVE FOR A RAPID (MAY BE DISCONTINUOUS) INCREASE OF VELOCITY WITH DEPTH (adapted from Reference 5).

¹In all the discussion to follow (Sections 2.1.1.1. through 2.1.1.4.), it is important to realize that the amplitude-epicentral distance relation does not take account of the effect of the earth's free surface. The present discussion is concerned with the amplitude of the wave incident at the earth's surface, whereas the motion of the free surface (the motion recorded by a seismograph) is a resultant of the displacements resulting from the incident wave and the reflected waves generated at the point of incidence. Thus there is a factor which can produce significant variations in the amplitude of the recorded earth motion resulting from the incidence of a body wave which is completely independent of the velocity-depth relation. Unfortunately, it is sometimes ignored in amplitude studies.

because of the presence of a shadow zone, i.e., a range of values of Δ for which no waves (except perhaps diffracted waves) arrive. The travel-time curve (see Figure 3) has a shadow zone between A and C which is followed by a double-branched curve. Large amplitudes are to be expected at the cusp at C.

2.1.1.4. Continuous Decrease in Velocity. In discussing the problem of the continuous decrease in velocity with depth, it is important to distinguish between those cases for which $\frac{r}{v} \frac{dv}{dr}$ either does or does not exceed the value of +1 in the region of interest, since no ray can bottom or have its deepest point at a level at which $\frac{r}{v} \frac{dv}{dr} \geq +1$.

Gutenberg [6] has considered the case of a discontinuous decrease in velocity gradient at a certain depth, followed by a constant-velocity gradient, and then a discontinuous increase in velocity gradient at some greater depth. Figure 4 shows the assumed velocity distribution, the ray paths, and the travel-time curves for various positions of the depth of focus. In this figure, the velocity is assumed to increase with depth to a value v_m , then to decrease as the depth increases, and finally to increase once more with depth. Furthermore, the assumed velocity variation is such that $\frac{r}{v} \frac{dv}{dr} \geq +1$ just below the level at which $v = v_m$. This level is designated by r_m . No rays can bottom in the region between r_m and the level below it at which r/v again equals r_m/v_m . This lower level is indicated by a dashed horizontal line in the portion of Figure 4 which shows the ray paths. The region in which no rays can bottom is called a "low-velocity channel." The limits and the width of the shadow zone depend upon the depth of focus, as may be seen either from the ray paths or from the travel-time curves. No shadow zone will exist if the focus is below the low-velocity channel. Large amplitudes are to be expected just beyond the shadow zone.

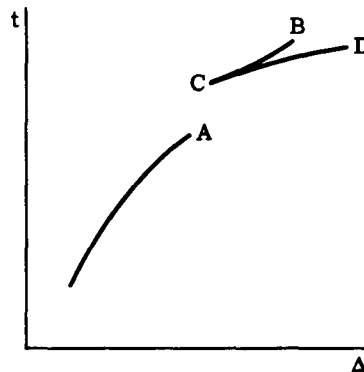


FIGURE 3. TRAVEL-TIME CURVE
FOR A DISCONTINUOUS DECREASE
OF VELOCITY WITH DEPTH
(adapted from Reference 5)

Båth [7] derived expressions for the shadow zone boundaries as a function of focal depth for an assumed crustal and upper-mantle velocity variation. He also derived expressions for the travel times and the energy of the emerging rays just before they strike the earth's surface.

If the decrease of v with increasing depth is slight enough that $\frac{r}{v} \frac{dv}{dr}$ remains less than +1, no low-velocity channel or shadow zone will exist. The travel-time curve is essentially the same as that shown in Figure 1, except for a possible reduction in curvature.

2.1.2. DETERMINATION OF BODY-WAVE VELOCITIES AT A GIVEN DEPTH. Two methods are available for determining the velocity of P or S waves as a function of depth within the earth. One is the classical Herglotz-Wiechert method; the other is a method derived by Gutenberg, employing deep-focus earthquake data.

It is assumed that the Herglotz-Wiechert method is so well known that it does not require a description here. This was the method used by Jeffreys [2] and Gutenberg and Richter [1] to determine the P- and S-wave velocity distribution in the mantle and the P-wave velocity distribution in the core. Since observationally determined travel-time curves are subjected to averaging, this method of interpretation as usually applied will be unable to detect any regional differences in the velocity-depth relation (assuming they exist), and will also be unable to detect any minor departures from a smoothed velocity-depth relation. The method formally fails if a level is encountered for which $\frac{r}{v} \frac{dv}{dr} \geq +1$.

Gutenberg's method [8] enables one to compute the velocity at the depth of focus. It requires a knowledge of the depth of focus and sufficient observations of the arrival time of the body wave at small epicentral distances so that one can determine the slope, $(dt/d\Delta)_i$, of the curve representing arrival time vs. epicentral distance curve at its point of inflection. If r_h and v_h refer to values of r and v at the focus, then

$$v_h = r_h / (dt/d\Delta)_i \quad (1)$$

for Δ measured in radians.

The ray arriving at the point of inflection of the travel-time curve is the one which left the focus horizontally. Figure 4 illustrates that no ray would arrive at the point of inflection if the focus were in a low-velocity channel, except for diffracted rays or possibly movements at the surface associated with a channel wave propagated within a few wavelengths of the surface. Gutenberg [9] states, "In such instances, a continuous travel-time curve is frequently drawn on the basis of all reported time, although these may refer to various types of waves. If the apparent velocity at the point of inflection of such a spurious travel-time curve is used, Equation 1 leads to a result which still is close to the actual velocity at the depth of the source. If, however,

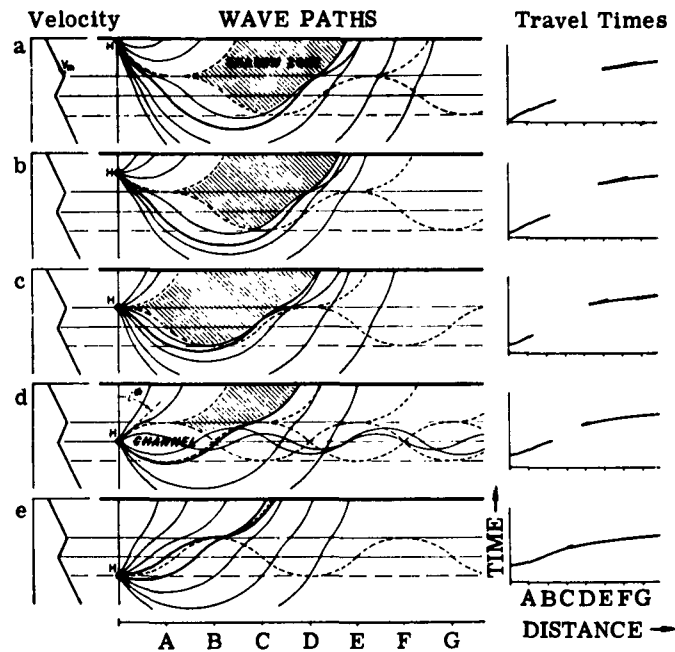


FIGURE 4. RAY PATH AND TRAVEL-TIME RELATIONSHIPS FOR A REGION OF CONTINUOUSLY DECREASING VELOCITY FOR VARIOUS DEPTHS OF FOCUS [6]

such a fictitious travel-time curve is used to calculate the wave velocity as a function of depth (using the Herglotz-Wiechert method), assumptions may have to be made, including discontinuities (as in the case of the '20° discontinuity') and a spurious velocity-depth curve may be obtained."

Figure 4 shows that Gutenberg's method is capable of producing the most serious errors in the determination of v for foci at or above the top of the low-velocity channel. For such focal depths, the point of inflection will be found on the left segment of the travel-time curve, and if there are insufficient data for distances just beyond the point of inflection, one would probably use later arrivals in determining the slope. This would result in too large a value for the slope $(dt/d\Delta)_i$, and consequently too small a value for v_h . Thus some caution must be exercised, both in applying this method and in using the conclusions based upon it, if a low-velocity channel is suspected to exist in the region of interest.

2.1.3. THE USE OF REFLECTED WAVES. The use of reflected waves has one great advantage over that of refracted waves for determining the depth to discontinuity surfaces, because waves can be reflected from a discontinuity involving either an increase or decrease in velocity; consequently, a low-velocity channel presents no difficulties. For the earth's upper

mantle, however, there is the problem that any discontinuities in velocity or velocity gradient may not involve a large enough contrast in the elastic parameters and the density in a sufficiently small depth interval to produce reflected waves of large enough amplitude to be identifiable. There is also a lesser problem of knowing the velocity-depth relation above the discontinuity in order to determine depth accurately. Seismological knowledge of the structure of the upper mantle, however, has not yet advanced to the point where the latter problem can be considered too serious.

2.1.4. THE USE OF WAVE PERIODS. Insofar as the upper mantle is concerned, studies of the periods of the body waves have been concerned primarily with identifying different branches of the travel-time curves. It has been observed that the periods remain essentially constant for a given branch, but change from one branch or segment to another.

2.1.5. THE USE OF TIME DIFFERENCES BETWEEN OBSERVED WAVES. Vesanen, Nurmia, and Porkka [10] have observed, using the Jeffreys-Bullen tables [11], that a linear relationship exists between the time differences $[(S - P)_{\text{surf.}} - (S - P)_h]$ and $(pP - P)_h$, where the subscript surf. refers to values for a surface focus, and the subscript h refers to values for a given focal depth. This relationship is practically independent of the epicentral distance. To apply the method, earthquakes in a given geographic region are selected, and from the seismograms $(S - P)_h$ and $(pP - P)_h$ are obtained. Only data in the epicentral distance range of 40° to 83° are used; this appears to be a good choice since in this interval of distances the S phase as seen on the seismograms is uncomplicated [12]. In practice the graph of $[(S - P)_{\text{surf.}} - (S - P)_h]$ vs. $(pP - P)_h$, using observed values of $(S - P)_h$ and $(pP - P)_h$, usually agrees with the curve obtained from the travel-time tables only for earthquakes having their foci below a certain depth; this critical depth depends upon the geographic region. They interpret the departures from the ideal curve to be the result of a difficulty in the analysis of the seismograms caused by the presence of a low-velocity channel and suggest that the depth to the level of minimum velocity in this channel is related to the value of $pP - P$ which separates the points that lie on and depart from the theoretical curve.

2.2. RESULTS FROM P-WAVE STUDIES

Seismologists do not completely agree upon the details of the structure of the earth's upper mantle. One must ask if these differences of opinion arise solely on the basis of the various interpretations, or if they are actually related to existing differences within the earth.

The latter point should be considered, because most investigators confine themselves to a limited set of data obtained from one or a few geographic regions. Furthermore, differences are known to exist in the structure of the earth's crust, and it does not seem unreasonable that

they could extend also into the mantle. For these reasons, the results from P-wave studies shall be broken down according to geographic area.

2.2.1. NORTH AMERICA. One of the important results of Byerly's study [13] of the Montana earthquake of 28 June 1925 was the observation that the slope of the travel-time curve of the P wave changed abruptly at an epicentral distance of 20° , with a decrease in slope at the larger distances. This discontinuity in slope of the travel-time curve was later found in other regions, and is now commonly referred to as the " 20° discontinuity."

Dahm [14] constructed travel-time curves of P and S for the Long Beach, California, earthquake of 1933, and computed the velocities of these waves in the mantle by applying the Herglotz-Wiechert method. His P-wave travel-time curve was practically a straight line to an epicentral distance of 15° , which implied an almost constant P-wave velocity in the upper 100 km of the mantle. He observed a sharp change in slope of the velocity-depth curve at a depth of 225 km, and found evidence of waves which could have been reflected from that depth. He also noted that the travel-time curve of a Japanese earthquake showed continuous curvature from 3° onward, and concluded that the subcrustal material is probably not horizontally homogeneous over the whole earth.

Gutenberg [9] has summarized the results of the studies of Gutenberg and Richter, who used the P waves of Southern California earthquakes. They observed that the P_n wave (with a speed of between 8.1 and 8.2 km/sec) was found at distances of 1° to 7° , with its amplitude rapidly decreasing. From about 7° to 13° the amplitudes of the waves were small and scattered about a straight line whose slope corresponded to an apparent velocity² of about 7.7 to 7.8 km/sec. Gutenberg considered this distance interval to be a shadow zone. At about 14° the shadow zone was terminated by the arrival of P waves of much larger amplitude and higher apparent surface velocity. These results are presented in Figure 5, in which the reduced travel times ($t - 13.6 \Delta$, where Δ is in degrees) and amplitudes are given as a function of distance. Gutenberg concluded, from the presence of the shadow zone, that a low-velocity channel (asthenosphere channel) exists. Figure 6 presents his view of the ray paths of the P waves for a focus in the crust, and the resulting travel-time curves.

DeBremaecker [15] also studied the variation of amplitude of the P wave with epicentral distance for earthquakes originating along the West Coast of North America and recorded

²By "apparent velocity" or "apparent speed" is meant the reciprocal of the slope of the travel-time curve. For a head wave traveling along a horizontal discontinuity surface, the true velocity equals the apparent velocity multiplied by the ratio of the distance from the center of the earth to the discontinuity surface to the radius of the earth.

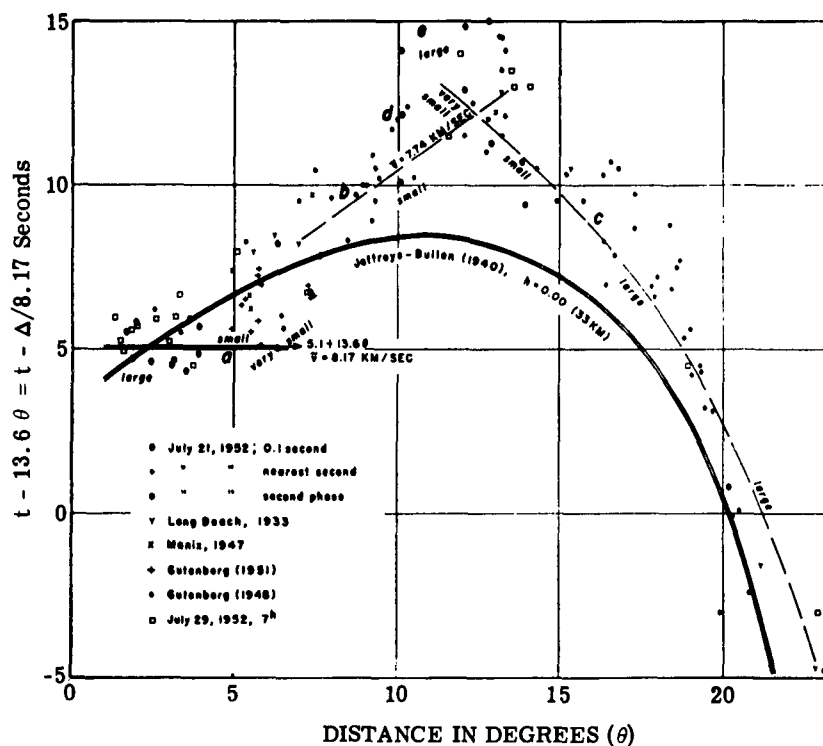


FIGURE 5. OBSERVED AMPLITUDES AND TRAVEL TIMES (REDUCED) OF P FOR SOUTHERN CALIFORNIA EARTHQUAKES [15a]

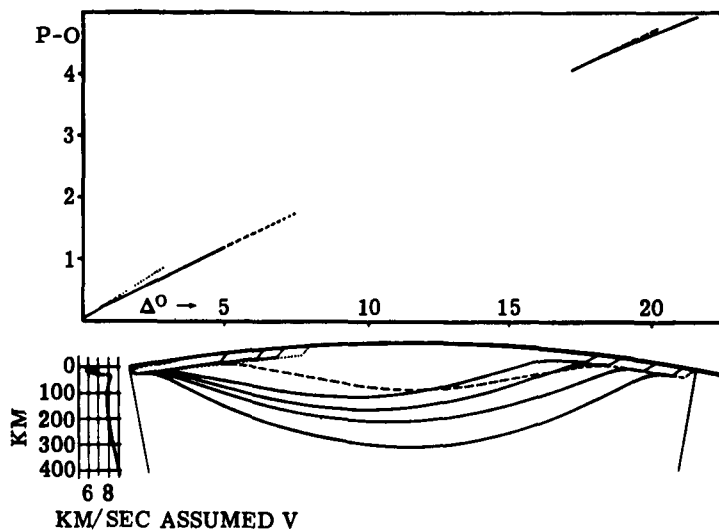


FIGURE 6. RAY PATHS AND CORRESPONDING TRAVEL TIMES OF P WAVES IN THE ASTHENOSPHERE CHANNEL [15a]

along the West Coast of the United States. Since all the stations and epicenters essentially lay on a line, there was no difficulty associated with the unequal azimuthal radiation of amplitude of earthquake waves, as long as an individual earthquake was considered. He found no coherence to the data when they were all grouped together and reduced to an equivalent magnitude according to the method of Gutenberg. However, the data from individual earthquakes showed a certain consistency, with the scatter of data confined to distances less than 16° . He concluded that the influence of regional factors is very important up to that distance and becomes smaller beyond it. His amplitude data, which applied to the maximum value of the horizontal component of the P motion, showed a minimum at 6° and a maximum at 14° , as did Gutenberg's. However, whereas Gutenberg had small amplitudes from 6° to 14° (the shadow zone), DeBremaecker's P amplitudes increased uniformly in this range. A possible explanation of this difference may result from DeBremaecker's use of maximum amplitudes and Gutenberg's use of the amplitude of the first wave. In any case, DeBremaecker concluded that, at a depth of 80 to 85 km, there is either a decrease in the rate of increase of velocity with depth or a decrease in velocity with depth, but not sufficient to produce a "low-velocity channel" and shadow zone.

Romney [16] in a study of the Dixie Valley-Fairview Peak earthquakes of Nevada, had 36 observations of P-wave travel times for distances less than 1200 km. He said the data could be explained in two ways: (1) a linear relation between travel time and distance from 300 km to at least 1200 km, with an apparent speed of 7.85 km/sec, or (2) parallel linear travel-time curves from 300 to 550 km and from 550 to 1200 km, with an apparent speed of 8.05 km/sec and with an offset of two seconds. Romney favored the latter interpretation, which he said could result from a 15-km increase in crustal thickness for the stations at the larger distances.

Ryall [17] noted that the travel times of the P waves from the Montana earthquake of 1959 showed scatter between 700 and 1400 km which could give the impression of being produced by a low-velocity channel in the upper mantle. However, he observed (see Figure 7) that in the region of scatter the arrival times tended to fall into groups according to physiographic regions — e.g., California stations, Pacific Northwest stations, Basin and Range stations—and that different groups indicate different apparent velocities. He stated that the travel times in the region of scatter could be explained on the basis of the known dip of the Mohorovicic discontinuity combined with regional differences in P_n velocity, and did not require a low-velocity channel in the mantle. To substantiate further his conclusion he noted that the amplitudes of the P waves did not decrease between 700 and 1400 km, except for the stations in the Pacific Northwest, which were found to be near a nodal line for P on the basis of a fault-plane solution.

Ryall found the slope of the travel-time curve to be fairly linear from 12° to 19° , with an apparent speed of 8.9 km/sec. Beyond 19° the travel-time curve began to vary continuously, but it showed a sharp decrease in slope at a distance slightly less than 25° .

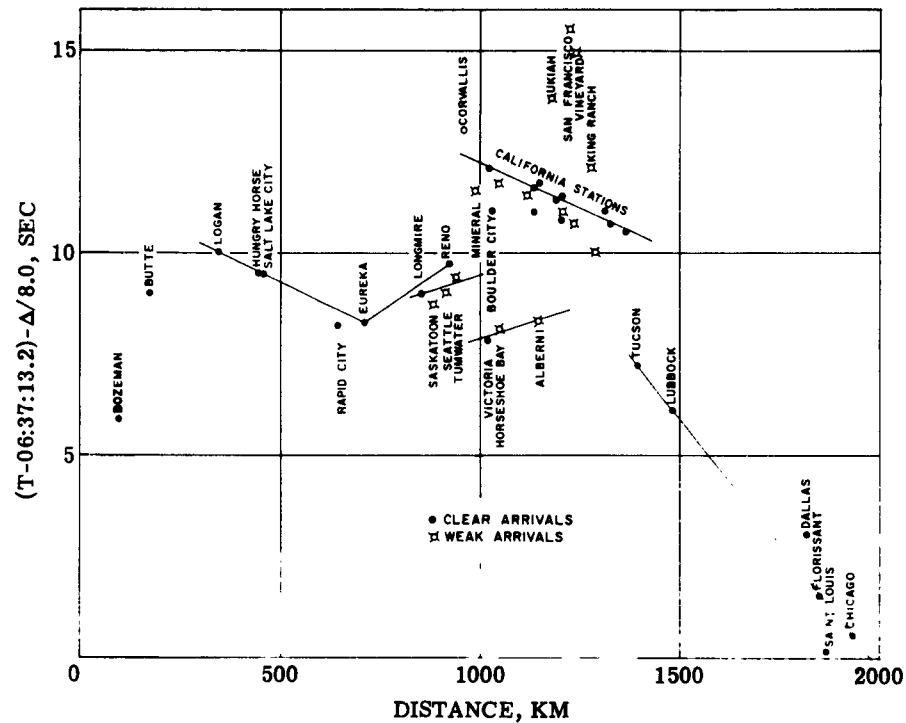


FIGURE 7. REDUCED TRAVEL TIMES OF P FOR EPICENTRAL DISTANCES LESS THAN 2000 km FOR 1959 MONTANA EARTHQUAKE [17]

Jeffreys [18] analyzed the travel times of P as reported in the International Seismological Summary for ten California and Pacific Northwest earthquakes which occurred during the years 1946 to 1952. He found that the mean slope of the travel-time curve between 2° and 10° was 13.95 ± 0.16 sec/deg, corresponding to an apparent velocity of 7.966 ± 0.091 km/sec.

The preceding discussion has primarily been concerned with the P waves of earthquakes occurring and recorded near the West Coast of North America. Lehmann [19] studied the travel times of P waves from five northeastern American earthquakes which occurred during the interval 1925 to 1940, and the 1952 Oklahoma earthquake. She found that the P curve could be taken as a straight line out to 14° , with a slope corresponding to a speed of 8.2 km/sec below the Mohorovicic discontinuity. Beyond 14° the data could be fitted to the Jeffreys-Bullen travel-time curve for surface focus which had been given a constant-time correction so that it would join the straight line curve at 14° . She found no existence of a shadow zone for P from these earthquakes.

Jeffreys [18] used the International Seismological Summary data of the same earthquakes studied by Lehmann and found the slope of the travel-time curve between 2° and 10° to be 13.59 ± 0.10 sec/deg, which gives an apparent velocity of 8.176 ± 0.060 km/sec.

Tatel and Tuve [20], in a study of P waves generated by a large chemical explosion in Tennessee, remarked that their results (including ten observations from 800 to 1600 km [7° to 15°]) did not show any shadow zones or delays. The P_n velocity was approximately 8.1 km/sec.

Reflections from discontinuities in the upper mantle were sought by Hoffman, Berg, and Cook [21], who studied the waves generated by large chemical explosions detonated near Salt Lake City, Utah. They found evidence of reflections from a discontinuity at a depth of 190 km, and some evidence for other discontinuities at depths of 520 and 910 km. They assumed, for the purpose of interpretation, flat-lying beds and an average speed of 7.62 km/sec from the surface of the earth down to the discontinuity surface.

Nuclear explosions have provided invaluable data for the study of the upper-mantle structure. Among the advantages of such a source are the accurate knowledge of the epicenter, the depth, and the origin time, and sufficient energy release to generate body waves which can travel to large distances. Another important advantage is a foreknowledge of the event, so that temporary seismograph stations can be set up to provide a much more complete set of data than is usually obtained from the recording of earthquakes. Bullen [22] and Griggs and Press [23] have described in some detail how nuclear explosions could provide unique data for some fundamental seismological research problems.

Even though most of the nuclear explosions at the Nevada Test Site were not conducted for the purpose of seismological experimentation, they have nevertheless provided some useful information to seismologists. Carder and Bailey [24] discussed the travel times of the P waves generated by explosions between 1951 and 1957 , including the underground RAINIER shot of September 19 , 1957 . Since most of their data for Nevada explosions were obtained at distances less than 6.5° , these data primarily furnished information concerning the crustal, but not the upper mantle, structure. On the basis of a somewhat limited amount of data, they concluded that the first arrivals of P would fit a straight-line curve, corresponding to an apparent surface speed of 7.95 km/sec, from 7° to 17.5° . They also observed a later arrival, with an apparent surface speed of 9.5 to 10.0 km/sec, which could be explained as a wave reflected from a discontinuity at a depth of somewhere between 160 and 185 km.

The underground nuclear explosions BLANCA and LOGAN, which took place in Nevada in October 1958 , provided a very comprehensive set of amplitude-distance data for P waves which had traveled through the upper mantle. Romney [25] discussed the seismological results obtained from these explosions. Portable stations, supplementing permanent stations, were established to form a line of stations extending east from Nevada to Arkansas, and then northeastward from Arkansas to Maine. The stations were approximately 100 km apart to a distance of 2300 km, and beyond that 250 km apart to a distance of 4000 km. Each of these stations was

equipped with a vertical-component Benioff, and most also had horizontal-component Benioffs oriented in directions along and perpendicular to the great-circle path at the station. The seismometer and galvanometer were loosely coupled, and the seismograph magnification was believed to be known to within 5% to 10%.

In his analysis, Romney used the maximum value of the ratio A/T (A = ground amplitude, T = period) in the first three cycles of the vertical component of motion. He found that at distances between approximately 200 and 1000 km, the amplitude (actually A/T) of P_n was inversely proportional to the cube of the distance, whereas P_n decreased to such small amplitudes beyond ~1000 km that it became undetectable at most stations. Other strong P waves appeared beyond 1000 km; their time of arrival was several seconds later than the expected time of P_n . These late waves generally had longer periods than P_n , and their travel times indicated a higher apparent surface speed. Romney said, "The abrupt change in the amplitude-distance relationship, the longer periods, the higher speeds, and a break in the travel-time curve all indicate the arrival of a new wave" [25]. The amplitudes of the P waves generated by these explosions agreed remarkably well with Gutenberg and Richter's data for Southern California.

Wright, Carpenter, and Savill [26], using the same seismograms as Romney, studied the relation of amplitude and period to distance for the P waves produced by BLANCA and LOGAN. They observed that the half-period of the first P arrival from 150 to 1000 km was practically constant. They also concluded that the decay of amplitude with distance in this range was greater than that predicted for a head wave, and stated that a non-frequency-dependent attenuation process was operative.

Werth, Herbst, and Springer [27] made a thorough study of the factors affecting the amplitudes of the vertical component of the surface motion to be expected in the P-wave group, and compared their results with observations obtained from the explosions BLANCA, LOGAN, and TAMALPAIS. They also used the same seismograms that Romney did. In deriving expressions for the amplitude, they considered not only the effect of head-wave attenuation along the discontinuity surface, but also the consequence of propagation through the crust and the presence of the free surface. On the basis of BLANCA and LOGAN data, they found experimentally that the effects of the free surface included a surface reflection or interaction factor of -3 in the range of 300 to 600 km. A surface interaction of -3 means that the free surface of the earth, by a reflection process, produces a motion which is three times as large, but opposite in sign, as the motion produced by the incident wave. For the BLANCA and LOGAN data, the surface interaction phenomenon is delayed by 0.4 seconds with respect to the incident wave, so that it gives the effect of producing a much larger amplitude to the second half-cycle of motion than the first half-cycle. On the basis of these investigators' work, it appears that it would be much more

reliable, in studying the variation of amplitude with distance in order to obtain information concerning the velocity variation with depth, to use only the amplitude of the first half-cycle of the first arrival, since this will be least affected by crustal structure. In fact, they obtained excellent agreement between the absolute value of the observed and predicted amplitudes for this first half-cycle of motion from 200 to 600 km, using a satisfactory value of a crustal model.

Specific studies of the travel times of the P waves resulting from BLANCA and LOGAN have been published recently. Before presenting these results, however, it would be of value to discuss briefly the results of the study of crustal structure in the vicinity of the Nevada Test Site. Press [28] used seismic refraction, Rayleigh-wave phase velocity, and gravity data to determine the crustal structure to the southwest of the Nevada Test Site. The seismic refraction results gave the following structure below the sediments: an upper layer 23 km thick with a P-wave velocity of 6.11 km/sec, and an intermediate layer 26 km thick with a P-wave velocity of 7.66 km/sec and a P_n velocity of 8.11 km/sec. P_n became the first arrival at distances beyond 320 km. In order to satisfy the phase velocity and gravity data, it was necessary to modify the thickness and/or the velocity (or Poisson's ratio) of the intermediate layer as determined by the refraction measurements. Diment, Stewart, and Roller [29] reported the results of seismic refraction and gravity measurements along a 300-km line southeast from the Nevada Test Site. A P wave with an apparent velocity of 6.15 km/sec was the first arrival for distances from 11 to 158 km. From 158 km to the end of the profile, the first arrival had an apparent velocity of 7.81 km/sec. The authors proposed a crustal model consisting of an upper layer 28.5 km thick, an intermediate layer 24.5 km thick, and a P_n velocity of 8.11 km/sec.

Adams et al. [30] reported the travel times of P waves from BLANCA and LOGAN to distances as great as 3850 km. They found that their data, out to 900 km, could be fitted by three straight-line travel-time curves. The apparent velocity to 128 km was 5.69 km/sec, from 128 to 481 km it was 7.65 km/sec, and from 481 to 900 km it was 8.12 km/sec. Assuming a horizontally layered crustal model, these data give thicknesses of 24 km for the upper layer and 36 km for the intermediate layer. The authors noted that consistent late arrivals for stations to the east of the Nevada Test Site indicated an eastward dip of the Mohorovicic discontinuity.

Lehmann [31] also analyzed the travel times of the P waves from BLANCA and LOGAN on the basis of data from temporary and permanent stations at distances of 60 to 4000 km. The travel times to distances of 325 km essentially satisfied the data of Press [28] for waves whose velocities were those of the upper and intermediate layers. P_n was observed as a first arrival from about 350 to 1600 km, with an apparent velocity of 8.08 km/sec. Beyond 1000 km its amplitude was very small, as noted by Romney [25]. As also noted by Romney, Lehmann found another strong P arrival beyond 1000 km which was several seconds later than P_n at a distance

of about 1000 km. She called this wave P_r . The extension of the P_n curve intersected the P_r curve at about 1700 km.

Lehmann constructed a simplified model of the crust and upper mantle which satisfied the observed arrival times. Its essential features are a practically constant P-wave velocity of 8.0 km/sec from the Mohorovicic discontinuity to a depth of 215 km, and a discontinuous increase in velocity at that depth to a value of 8.35 km/sec. She noted that the travel times could be better fitted by including a layer with a velocity of 7.90 km/sec at depths from 150 to 215 km, and remarked that such a layer does not seem unlikely but its existence is by no means proved. She stated that there is no low-velocity layer of any importance for P waves, as evidenced by the fact that the travel times of P_r are so small that the P_r curve intersects the P_n curve at a distance of only 1700 km.

The underground nuclear explosion GNOME, detonated in New Mexico on December 10, 1961, produced some interesting seismological results. Romney et al. [32] presented travel-time and amplitude data obtained from this event. Figure 8 shows the location of the seismograph stations in North America which recorded the elastic waves generated by the explosion. Most of the data on amplitudes used in this study were obtained from the 40 stations operated by the Geotechnical Corporation, which are indicated by large black circles in Figure 8. The data of all the observing points were used in the study of the travel times.

The travel-time curves to a distance of 400 km (which are strongly influenced by the crustal structure in the region of New Mexico) are reproduced as Figure 9. One can see from the figure that the speed of P_n , obtained from arrivals at distances between about 200 and 400 km, is 8.1 km/sec. Beyond 400 km the first arrivals are divided into two distinct groups: those which propagated to the east, and those which propagated to the west. Travel times to the east were significantly earlier than travel times to the west for comparable distances (see Figure 10). In this figure the reduced travel times ($t - \Delta/8.1$, where Δ is in km) of the first arrival are plotted for the distance range of about 200 to 2500 km. Figure 10 illustrates that from 200 to 400 km there is no difference in travel time to the east or west, and that the P_n speed is 8.1 km/sec. Between approximately 400 and 2000 km (the distance interval which shows the effect of the structure of the upper mantle), the apparent surface speeds of the first P arrivals to the east were higher than those to the west. The Jeffreys-Bullen travel-time curve for surface focus was included in this figure for comparison purposes. Although it fits neither set of data, it does represent a good average value, which is not surprising since it was determined statistically from large amounts of data from all over the world. A "contour" map, showing time departures from the Jeffreys-Bullen curve, is shown in Figure 11. From this it might be concluded that both the California-Nevada region and the Mississippi-Ohio River Valleys region are anomalous areas.

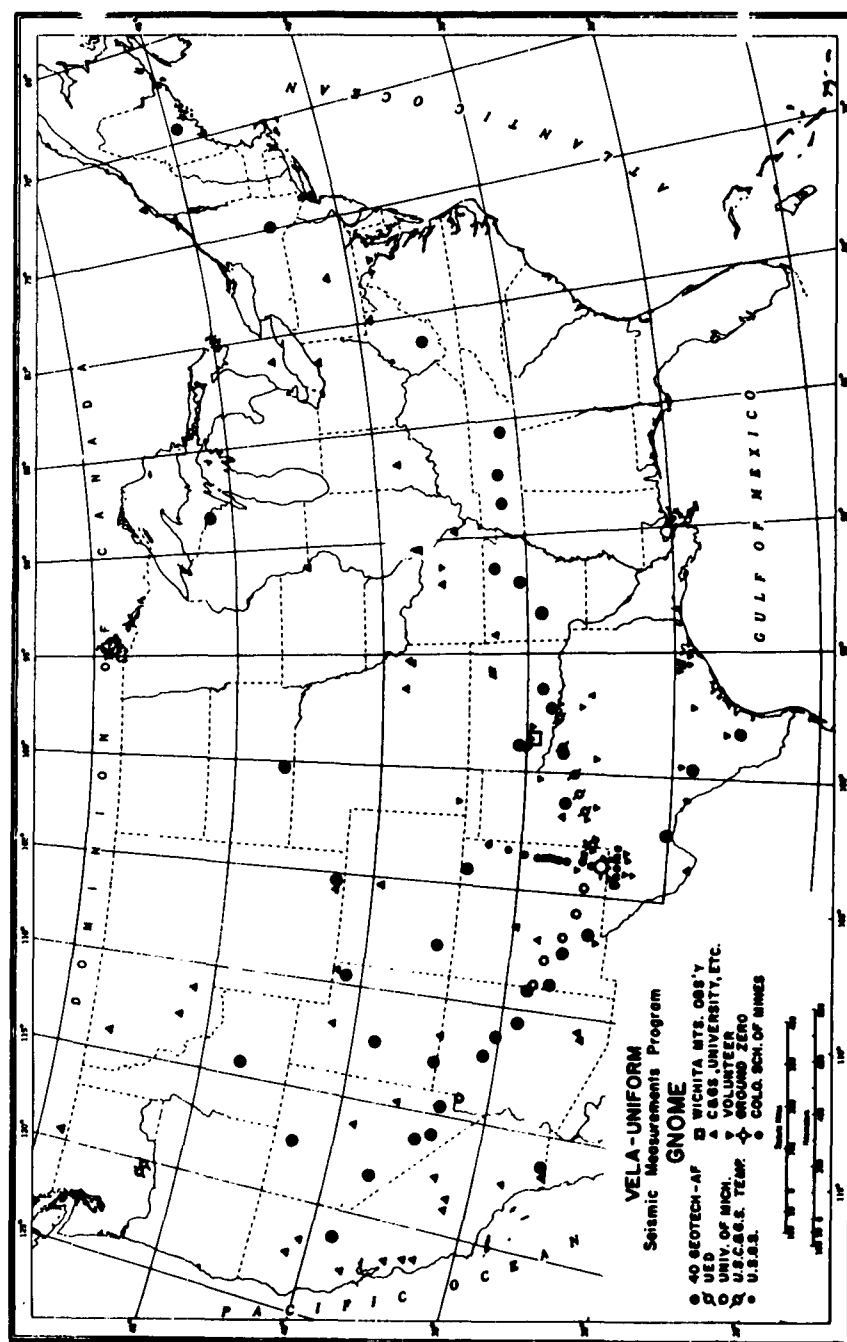


FIGURE 8. LOCATION OF SEISMOLOGICAL STATIONS IN THE UNITED STATES WHICH RECORDED THE GNOME EXPLOSION [32]

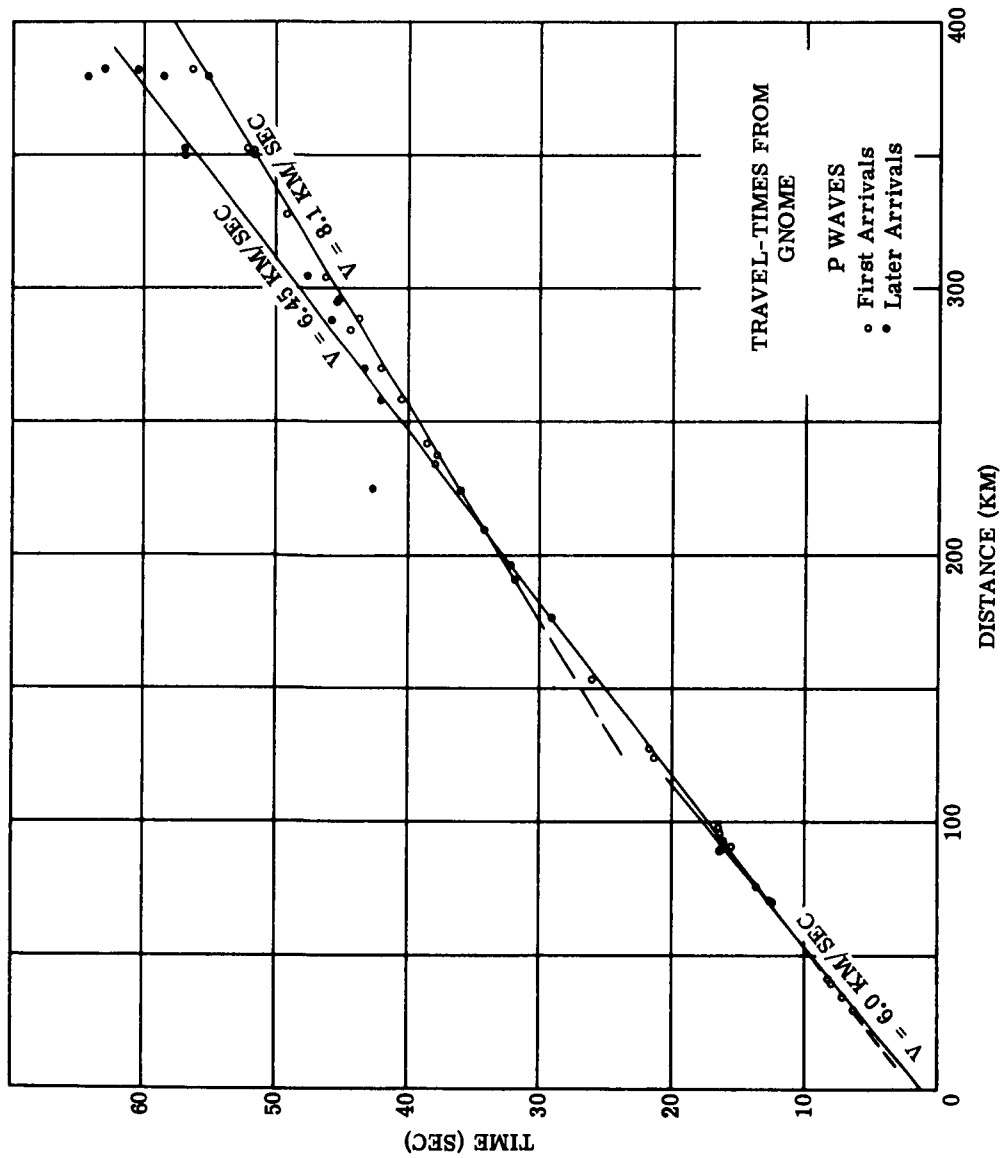


FIGURE 9. TRAVEL-TIME CURVES OF P WAVES FROM GNOME FOR SMALL EPICENTRAL DISTANCES [32]

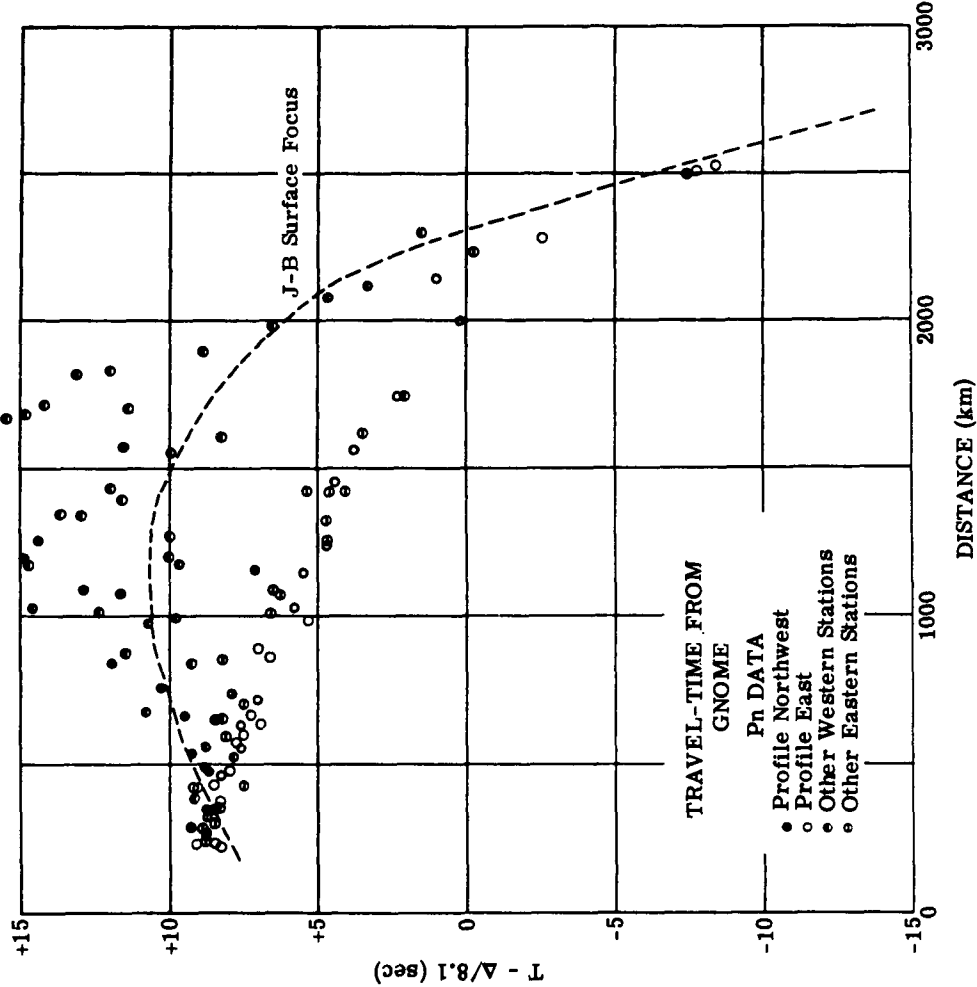


FIGURE 10. REDUCED TRAVEL TIMES OF P FROM GNOME [32]

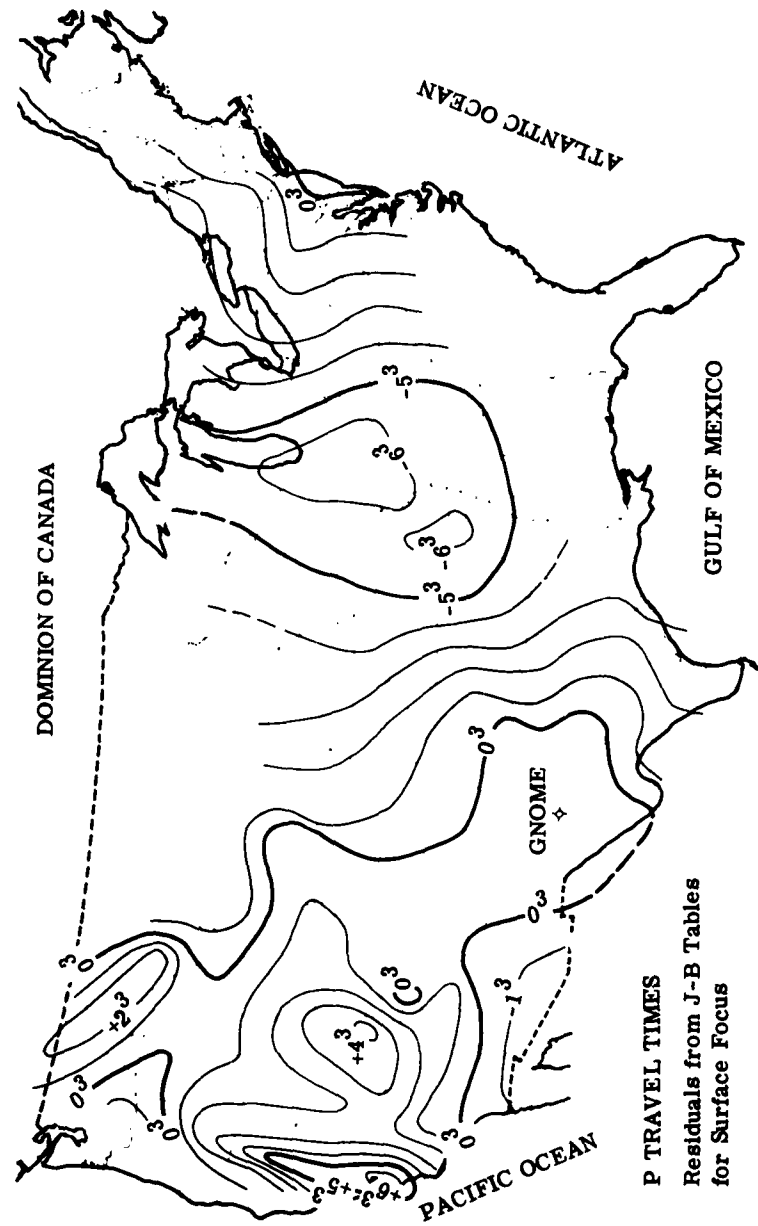


FIGURE 11. RESIDUALS OF TRAVEL TIMES OF P FROM GNOME, BASED ON JEFFREYS-BULLEN TABLES FOR SURFACE FOCUS [32]

The authors also observed that, however measured, the amplitudes of the first P arrivals to the east were clearly larger than to the west. The extreme variation in amplitudes occurred at a distance of 1500 km, where the motion at Jackson, Tennessee, was 150 times larger than at Mina, Nevada.

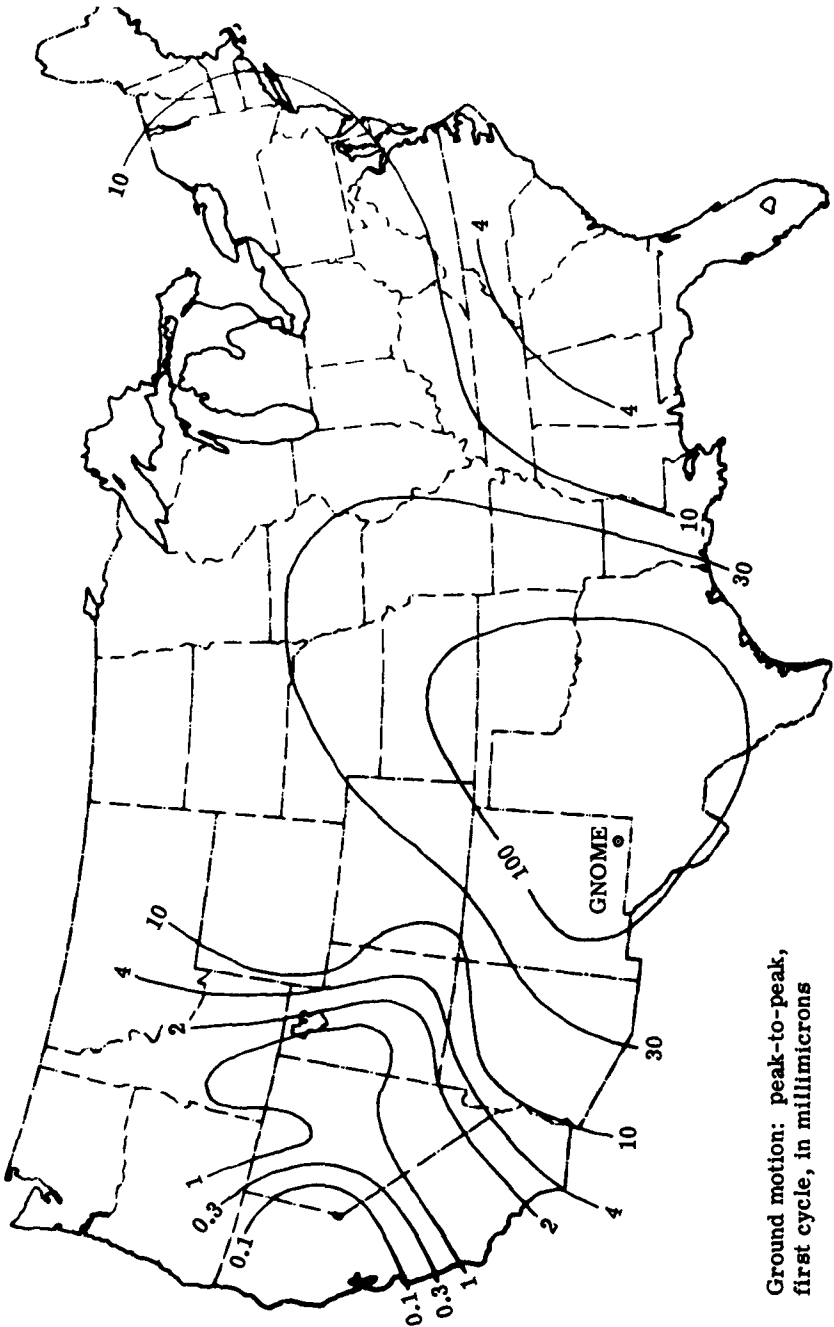
Herrin and Taggart [33] have combined studies of the amplitudes and travel times of the P waves recorded in North America from GNOME with studies of the travel times of nuclear explosions in Nevada and with recent earthquakes in Montana and southeastern Missouri. The results of the amplitude study of the GNOME P-wave data are shown in Figure 12, which shows clearly that the amplitude attenuation is much greater to the west and northwest than to the northeast.

Their studies of the travel times of waves generated by the nuclear explosions and earthquakes mentioned above demonstrate that there is a substantial variation in the regional velocity of P_n in the United States. They define P_n as the first arrival in the interval from the crossover distance at less than 200 km to a maximum distance of 2000 km.³ Herrin and Taggart determined P_n -interval velocities from adjacent pairs of stations, using explosion and earthquake data, and then plotted a P_n -velocity contour map. This map (see Figure 13) is interesting because of the wide range in P_n values (7.5 to 8.5 km/sec), of the unusually low values in the west, and the unusually high values in the South-central United States. The authors noted that P_n interval velocities in Arizona, Nevada, and Utah which were determined from explosions or earthquakes 1000 km away agreed closely in most cases with values found using refraction spreads up to 200 km in length. They therefore concluded that P_n appears to travel as a head wave to distances of 1000 km in the west. They remark that an assumption concerning a similar path for distances from 500 to 2000 km in the Eastern United States needs to be tested.

Herrin and Taggart note that large P_n amplitudes correlate with early arrival times, and small amplitudes with late arrival times. They suggest that the low-velocity layer in the upper mantle may begin at the base of the crust in areas where the P_n velocity is less than 7.8 km/sec, and that in the Eastern United States it may be deep or nonexistent.

A most significant result of their study is the demonstration that regional variation in P_n velocity can give rise to large errors in epicentral determination, if one uses data from stations less than 20° distant and assumes that the travel time is only a function of the epicentral distance. For the GNOME explosion, for example, the "epicenter" located in this way was 16 km east of its true location.

³ This differs from the more restricted meaning assigned to P_n throughout the remainder of this report, in which it is taken to be a wave which has traveled as a head wave along the Mohorovicic discontinuity.



Ground motion: peak-to-peak,
first cycle, in millimicrons

FIGURE 12. AMPLITUDES OF FIRST CYCLE OF P MOTION FROM GNOME [33]

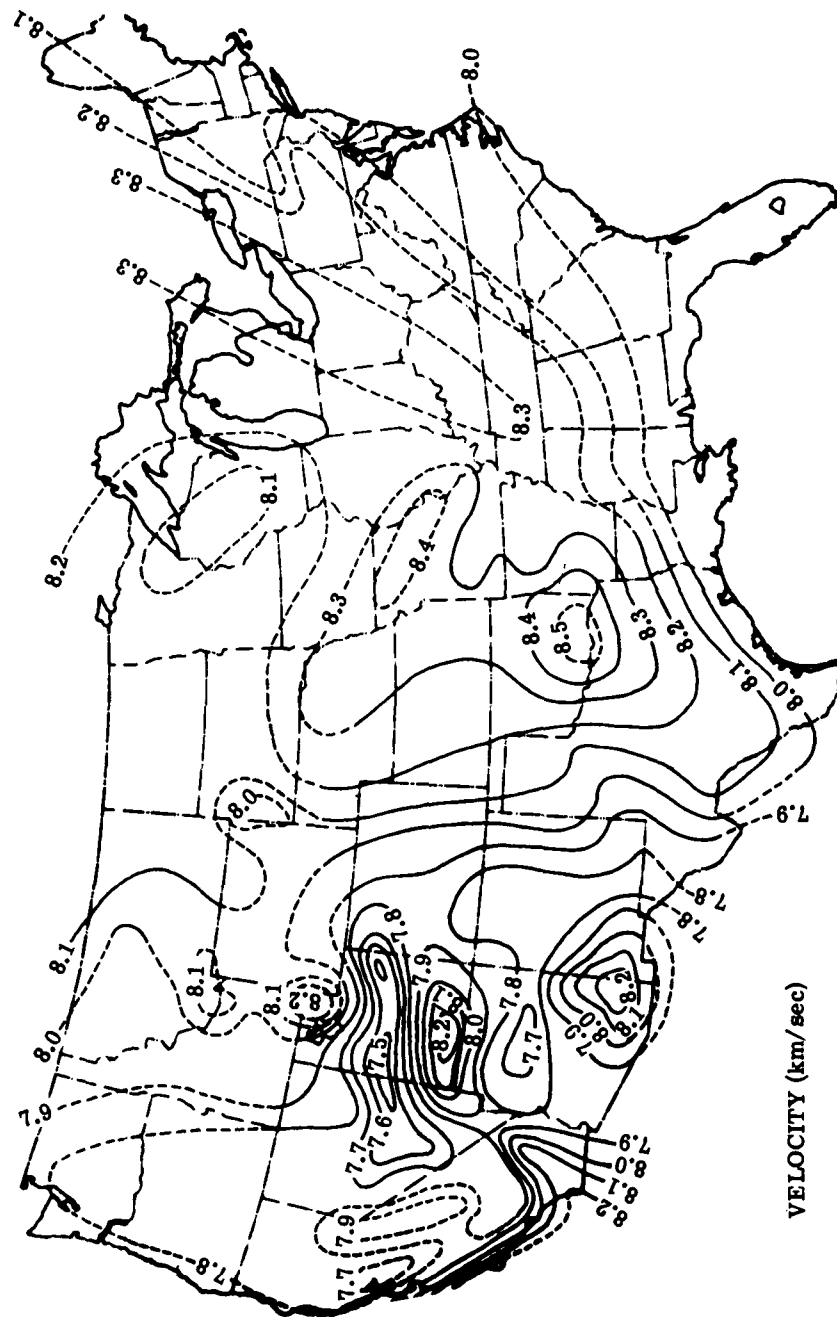


FIGURE 13. P_n VELOCITY MAP OF THE UNITED STATES [33]

2.2.2. EURASIA. The first seismological evidence for the existence of a low-velocity channel in the upper mantle was presented by Gutenberg [34], who studied the variation of amplitude with distance for two European earthquakes that occurred in 1911 and 1913, and for a 1923 Japanese earthquake. The P waves of the European earthquakes showed a marked decrease from a distance of about 500 to 1700 km, and then increased rapidly at 1700 km.

An opposing view of the P-wave velocity variation in the upper mantle has been held by Jeffreys. In a recent work concerning Europe [35], he incorporated his earlier investigations with the results of his studies of 24 earthquakes which occurred during 1946 to 1949. He found the travel-time curve of the P wave to be practically linear to 14° , with an apparent speed of 8.13 km/sec. He found no evidence of a shadow zone for P between 8° and 14° , but rather of a triplication of the travel-time curve (see Figure 2) at somewhat larger distances with the cusps near 14° and 20° . This would indicate that there is either a rapid or discontinuous increase of velocity with depth in the upper mantle.

Jeffreys' [18] latest work substantiates his earlier findings. The slope of the travel-time curve and the apparent velocity of P for short distances are 13.73 ± 0.13 sec/deg and 8.093 ± 0.077 km/sec, respectively. Using mean values for the 2° to 10° distance interval, he obtained values of 13.66 ± 0.07 sec/deg and 8.140 ± 0.041 km/sec.

Willmore's [36] study of the Heligoland (northwest Germany) explosion showed that the P waves were recorded to a distance of about 1000 km, with no suggestion of a shadow zone. Their travel times fit a curve with a slope of 13.6 sec/deg (apparent speed of 8.18 km/sec).

Lehmann [37] constructed a velocity-depth curve for P in the upper mantle which was made to satisfy a mean P travel-time curve for Europe. The travel-time curve, which was obtained by Jeffreys [38] from a statistical study of a large number of earthquakes and observations, was essentially the same as that mentioned above (Jeffreys, [35]). In order to explain the bend in the travel-time curve near 14° or 15° , Lehmann pointed out that either there must be a strong increase in velocity at a depth not much greater than 120 km (there is evidence to show this is not the case), or else the travel-time curve from 15° onward is not the continuation of the curve at smaller distances, but rather a different branch of the curve caused by a wave refracted in a lower layer. She accepted the latter possibility and tentatively placed the depth at which the velocity rapidly increases at 220 km. In order to explain the relatively large P amplitudes to about 5° followed by decreasing amplitudes to 15° , she assumed an increase in the P velocity below the Mohorovicic discontinuity from 8.0 to 8.12 km/sec in a depth interval of 20 km, and then kept the velocity constant to a depth of 220 km. The velocity was assumed to increase abruptly at this depth from a value of 8.12 to 8.4 km/sec, with a strong increasing gradient below. Insofar as the travel times are concerned, it would make little difference if the region of constant

velocity were replaced by one in which the velocity either increased slightly or decreased slightly with depth.

Using the same velocity-depth relation, Lehmann computed P-wave travel-time curves for focal depths of 130 and 285 km, and compared these with observed times of European earthquakes at those depths. The observed and computed travel times compared satisfactorily. However, she noted that her velocity-depth relation did not provide a unique solution.

Commenting further upon the uniqueness problem, Lehmann [39] noted that the increase in velocity at 220 km must be discontinuous, because there are no large amplitudes at the cusp in the travel-time curve. She noted that it was possible to have a slight decrease of P velocity in the depth interval of 150 to 220 km, although the P-travel times do not require such a decrease of velocity.

Gutenberg [9] applied his method of determining the velocity at the depth of focus to the P-wave travel times of 25 earthquakes which occurred in Rumania and Hindu Kush. From this analysis he found a velocity minimum of about 7.8 km/sec at 75 km depth, which would be sufficient to produce a "low-velocity channel" beginning at the Mohorovicic discontinuity.

Ruprechtová [40] observed (from a study of the travel times of P waves recorded at Prague in the epicentral distance range of 10° to 30°) that the largest curvature of the travel-time curve occurred at 17° to 18° . Associated with this large curvature was a maximum value of P-wave amplitudes. In a later paper [41], Ruprechtová noted from an additional study of five shallow focus earthquakes recorded at 35 European stations that the amplitude decreased rapidly from a maximum at 2° to a minimum at 11° , and then increased gradually to a maximum value at 19° .

Jeffreys [18], using the travel times of eleven earthquakes in central Asia for the years 1947 to 1952, found that the P travel-time curve to a distance of 30° was the same for this region as that he found for Europe (1958). The slope of the travel-time curve between 2° and 10° was 13.64 ± 0.10 sec/deg, with a corresponding apparent velocity of 8.146 ± 0.060 km/sec.

Shirokova [42] studied the amplitudes of the P waves of seven earthquakes which occurred in Hindu Kush at a depth of 200 km and which were recorded by fifteen Russian seismograph stations. In order to eliminate the effect of the variation of amplitude with azimuth of the waves radiating from the focus, she first determined the focal mechanism using P- and S-wave data, and then normalized the data as if the source were a uniform one. These normalized data indicated a shadow zone at distances between 500 and 700 km, which implies that the "low-velocity channel" must extend at least to the depth of focus, namely 200 km. She determined

a model of the crust and upper mantle (see Figure 14) that satisfied both the observed travel times and amplitudes of P. Although the model is not unique, she believes that the velocity gradient at the top of the mantle must be large (from 8.0 to 8.7 km/sec) and that the boundaries of the low velocity layer (7.8 km/sec) appear to be discontinuous.

Tryggvason [43] studied the P waves of four earthquakes in the Arctic-Atlantic ocean region. The locations of the earthquake epicenters, together with the year of their occurrence, are indicated in Figure 15. The dotted region in this figure corresponds to the seismic belt associated with the mid-Atlantic Ridge. Tryggvason found, from the analysis of the P-wave travel times, that an unusual upper mantle structure underlies the mid-Atlantic Ridge and the adjacent area. His results are summarized in Figure 16. He found the P_n velocity in Northwest Europe and Scandinavia to be 8.36 km/sec, with no evidence of a shadow zone. Beneath the oceanic ridge and seismic belt his model contains a broad zone of low-velocity material in the upper mantle, with a P_n velocity of 7.4 km/sec, extending to a depth of 140 km, and then increasing discontinuously to 8.2 km/sec. Neither the velocity-depth relation nor the horizontal limits of the 7.4 km/sec material are uniquely determined, but Tryggvason gives a probable width of 1000 km for the extent of the upper mantle low-velocity region and says that it may possibly cover the whole oceanic region between Greenland and Norway. He is careful to point out that the structure which is classified here as oceanic is different from that generally found below the deep oceans.

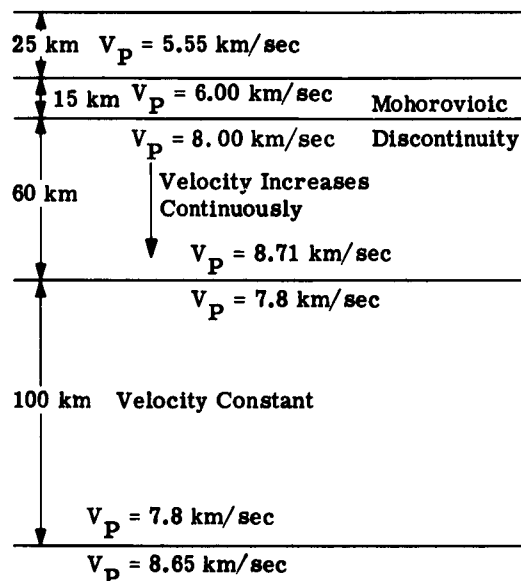


FIGURE 14. SHIROKOVA'S MODEL OF THE P-WAVE VELOCITY VARIATION IN THE CRUST AND UPPER MANTLE (adapted from Reference 42).

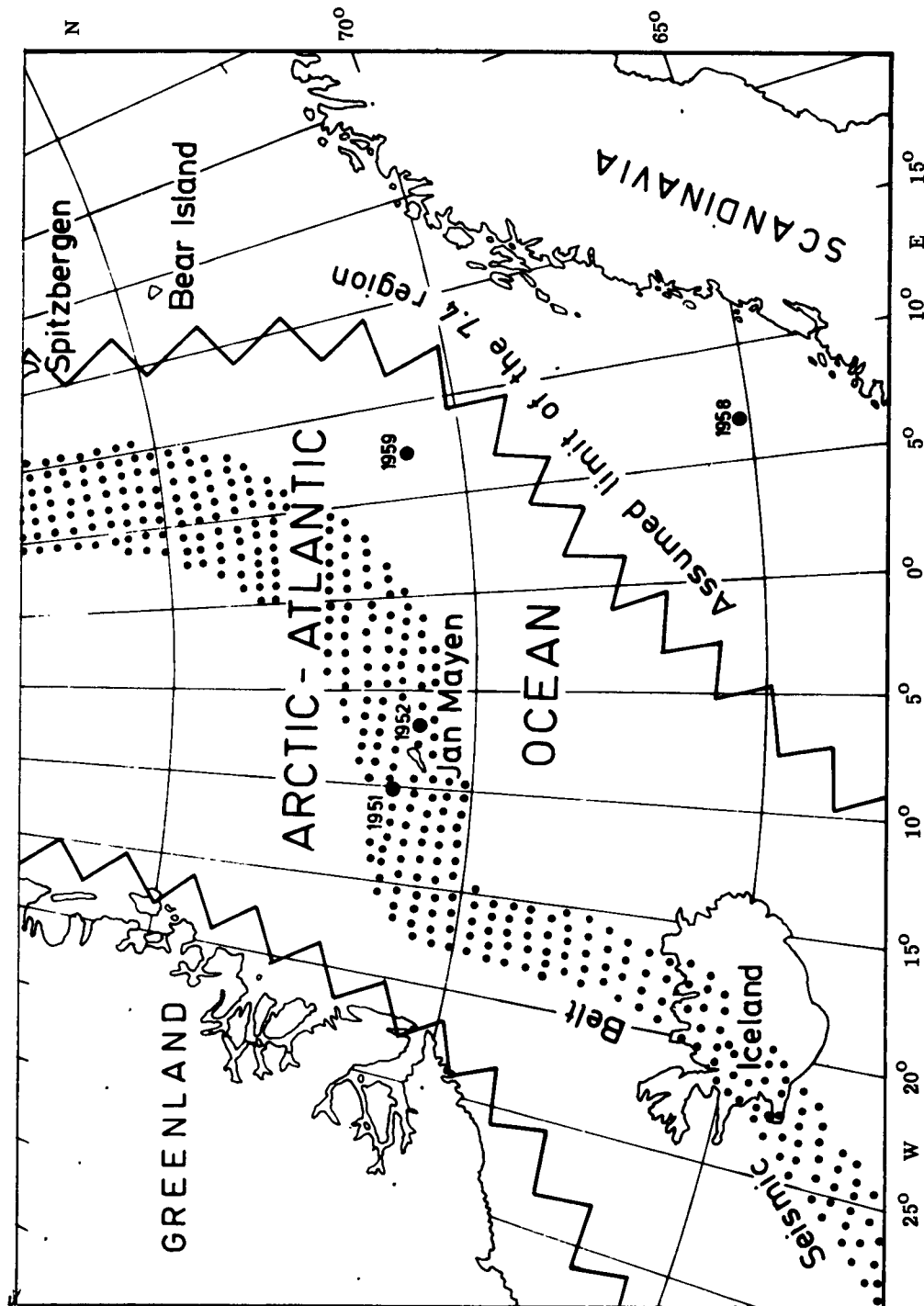


FIGURE 15. LOCATION OF EPICENTERS IN THE ARCTIC-ATLANTIC OCEAN [43]

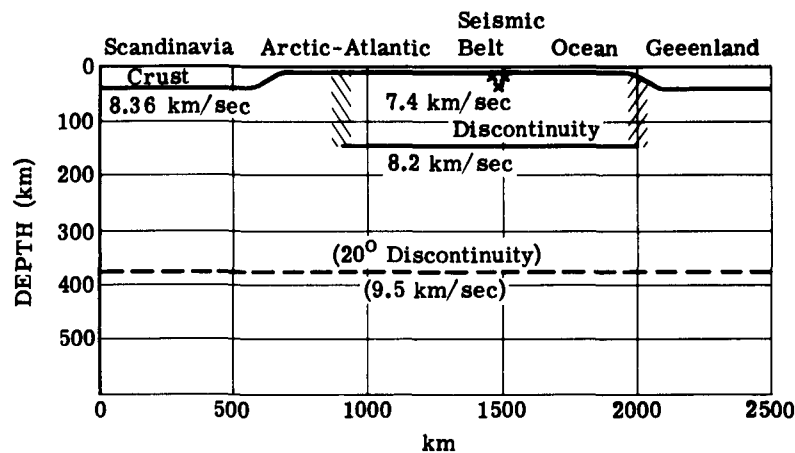


FIGURE 16. UPPER MANTLE STRUCTURE AT SCANDINAVIA, ARCTIC-ATLANTIC OCEAN, AND GREENLAND [42]

Ewing and Ewing [44] made explosion refraction studies in the same oceanic area as was studied by Tryggvason. In the western and eastern basins of the North Atlantic Ocean they found P_n velocities of 8 km/sec and 7.7 to 7.8 km/sec, respectively. The mid-Atlantic Ridge area and the Norwegian and Greenland Seas, however, showed only two refracting layers, with velocities averaging 5.6 and 7.4 km/sec. They considered the ridge structure to be about 700 km wide. Although they recorded no seismic velocities higher than 7.5 km/sec in this region, they assumed that the 7.4 km/sec material was of the order of 25 km thick, and that below it the velocity would increase to 8.0 km/sec. However, Tryggvason's P arrivals require a much larger thickness of the low velocity upper mantle material, since there are time delays to the west as compared to the east of the order of eight seconds in the distance interval of 10° to 15° .

2.2.3. JAPAN. Most of the earthquakes for which Gutenberg applied his method of determining the velocity at the depth of focus occurred in the region of Japan, the Marianas, and the Kurile Islands. The results of these studies are summarized in Figure 17 (Gutenberg [9]). In this figure, V and v refer to the velocities of P and S waves, respectively. The dash-dot lines refer to velocities obtained from independent studies of shallow earthquake and explosion generated waves. The slope of the line with the arrowhead indicates the minimum value of the magnitude of the velocity gradient required to produce a low-velocity channel. Figure 17 shows that the minimum value is already exceeded just below the Mohorovicic discontinuity (the beginning of the dash-dot curve).

Jeffreys [38] revised his travel-time curve for P waves in Europe and made an additional study of the travel times in Japan. He found no evidence for a change from his original travel

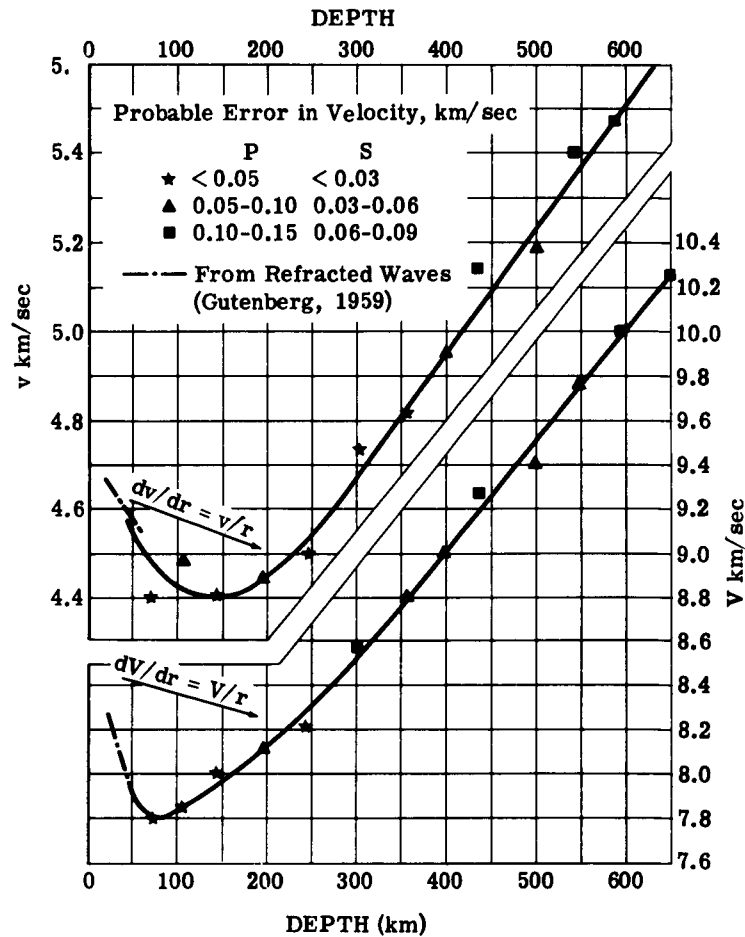


FIGURE 17. VELOCITIES OF P(V) AND S(v) WAVES AS DETERMINED FROM DEEP-FOCUS EARTHQUAKES [9]. The earthquakes occurred principally in the region of Japan, the Mariannas, and the Kurile Islands.

times [2] for earthquakes occurring and recorded at or near Japan. Thus he concluded that there is a regional difference between Europe and Japan which can produce differences in travel times for P to 20° by as much as four seconds. In his latest work [18], Jeffreys reiterates his previous conclusions. He found a value of 14.13 ± 0.04 sec/deg for the slope of the travel-time curve from 2° to 10° , which results in an apparent velocity of 7.870 ± 0.024 km/sec. This value is significantly lower than that found for continental regions, such as Europe, central Asia, and Eastern North America. He found the slope of the travel-time curve was nearly constant out to 15° .

An excellent summary of the research performed at the University of Kyoto on the subject of travel times and upper-mantle velocity distribution has been given by Nishimura, Kishimoto, and Kamitsuki [45]. They made a study of the travel times, amplitudes, and periods of P waves recorded at epicentral distances of 6° to 26° by 23 stations of the Japan Meteorological Agency, all equipped with similar Wiechert seismographs. The purpose of their study was to decide between Gutenberg's and Jeffreys' model of the upper-mantle structure beneath Japan. They observed that the travel-time curve of first arrivals showed an abrupt change in slope near 20° , as well as a change in amplitude and wave period. Moreover, they could follow the branch which appeared as a first arrival beyond 20° back to distances of 13° on the basis of wave periods. All of this agrees with Jeffreys' version of the travel-time and velocity variation of P [2]. Thus, Nishimura et al. found evidence for the existence of the 20° discontinuity in Japan, but their data did not permit them to decide whether it corresponded to a discontinuous change in velocity or in velocity gradient in the upper mantle. They looked unsuccessfully for a decrease in amplitude between 5° and 15° , as called for by Gutenberg's model. Except for a single point, one could conclude that their data fitted Jeffreys' model and not Gutenberg's. This point concerns the fact that they found P_n velocities of 8.1 km/sec, whereas Jeffreys' model calls for such velocities to be 7.8 km/sec. Thus they concluded that there must be some region near the top of the mantle in which the velocity decreases with depth, but in a manner different from that proposed by Gutenberg, and that there is a greater depth (associated with the 20° discontinuity) at which the velocity or velocity gradient increases discontinuously.

2.2.4. OTHER AREAS. Gutenberg and Richter [46] studied South American earthquakes recorded at Huancayo, Peru, at distances between 2° and 23° , and focal depths between 50 and 250 km. They found that a shadow zone existed for earthquakes of depth less than 200 km, but disappeared for greater depths. The shadow zone had its greatest width when the focal depth was about 80 km. Thus they concluded that there is a slight decrease in P-wave velocity at a depth of approximately 80 km.

Vesanen, Nurmi, and Porkka [10] applied their method of analysis, described earlier in this report, to a number of earthquakes in different geographic regions. On this basis they found depths to the minimum-velocity level of 45 km for Alaska, 80 km for north Japan, 95 km for the Tonga area, and 120 km for South America.

Bolt, Doyle, and Sutton [47] studied the P wave produced by the 1956 atomic explosions in Australia, which were recorded at distances between 0.4° and 11° by stations set up especially for this purpose. They found that the first arrivals were uniformly well recorded over the entire distance range, with no evidence of a shadow zone. Their travel times indicated a P_n velocity of

8.21 km/sec. They noted the similarity of this P_n velocity to that of other shield areas, namely the Canadian Shield (8.18 km/sec) and the western Transvaal of South Africa (8.27 km/sec).

Cleary and Doyle [48] analyzed the travel times of P for an earthquake which occurred on May 21, 1961 about 60 miles southwest of Sydney, Australia. The first arrivals could be fitted by a straight line out to 14.9° , with an apparent velocity of 8.16 ± 0.03 km/sec. They noted that the P-wave travel-time curve for Eastern Australia is linear out to at least 15° , with some possible attenuation in amplitudes between 8° and 10° .

Rocard [49] has described the amplitudes of P waves resulting from a 60- to 70-kiloton French nuclear explosion in the Sahara. He noted that there was no shadow zone in the distance range of about 500 to 1500 km, corresponding to P waves recorded in Europe. Commenting on the results of the United States nuclear explosions in Nevada and New Mexico, he remarked that they indicate that the Mohorovicic discontinuity surface is regular to the east of New Mexico, but irregular to the west. He calls upon focusing effects produced by the curvature of the discontinuity surface to account for abnormally small or large P wave amplitudes, and thus does not associate them with any abnormal velocity variation with depth in the upper mantle.

Finally, a reference should be made to Gutenberg's work [50] in which he tabulated the values of P_n velocities found by investigators throughout the world during the 1950's. All the values given lie between 7.91 and 8.32 km/sec. He stated that these data establish the fact that the P_n velocity decreases as the crustal thickness increases, and he derived a formula relating these quantities on the assumption that the relation is linear.

2.3 RESULTS FROM S-WAVE STUDIES

Observations of S waves at distances less than 20° are much more limited than those of P waves. There are several reasons for this scarcity of S-wave data. The one most commonly given is that the beginning of the S motion may be confused by later motion in the P-wave group. That this is a real problem may be seen by an inspection of Figure 18, in which the event noted by F is Byerly's [51] false S phase. The true S_n phase, which was identified on the basis of an increase in frequency, arrived approximately 20 seconds after F on the seismograms shown in Figure 18. Byerly observed F from earthquakes originating in California and recorded at distances between 2° and 9° . He found that its travel-time curve could be represented by a straight line whose slope gave an apparent speed of 4.35 to 4.5 km/sec, the same value as for S_n . He also observed that the horizontal and vertical components of motion were in phase and corresponded to the motion of a longitudinal wave. In a more recent study, Cameron [52] reported an apparent speed of 5.1 km/sec for the false S wave. He also noted that its particle motion corresponded to that of a longitudinal wave, and in particular one arriving along a nearly horizontal ray path. He

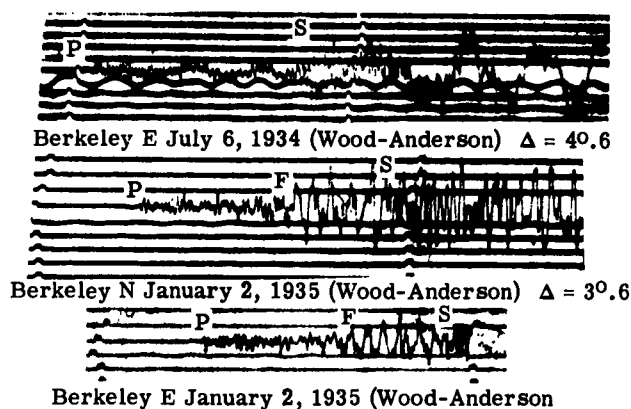


FIGURE 18. SEISMOGRAMS SHOWING THE FALSE S PHASE [51]

found it present only in shallow-focus earthquakes occurring in a region of continental structure, and identified it as a wave traveling a direct path from the focus to the station. Oliver and Major [53] suggested that the false S wave may be explained as a leaking-mode wave.

Another reason why S-wave data are not plentiful in this distance range may be that most of the high-magnification seismographs in operation have a rather selective frequency response which favors the recording of P waves and discriminates against the recording of the somewhat lower frequency S waves.

A third reason for the difficulty in obtaining adequate S-wave data at distances less than 20° is the effect of the earth's free surface on the recorded S-wave motion. The motion recorded by the seismograph is not the motion of the S wave as it approaches the station, but rather the resultant of this motion and that of the P and S waves reflected at the free surface. If the medium is not layered, the SH motion of the surface is always twice that of the SH component of the incident wave. However, the SV component of the surface motion varies in a complicated manner with the SV motion of the incident wave, depending on the angle of incidence at the surface (see Nuttli [54]). For angles of incidence at the surface greater than $\sim 35^\circ$, the SV surface motion will be elliptical, the sense of rotation being retrograde for angles of incidence between 35° and 45° , and prograde for angles of incidence greater than 45° . In addition, both the vertical and horizontal components of the recorded SV motion will usually be out of phase with respect to the SH motion. If one assumes that the S-wave velocity at the earth's surface is greater than or equal to 3.5 km/sec, it can be shown that the angle of incidence of S will be greater than 35° for all distances less than 20° . Thus one must expect a complicated surface-S motion for distances less than 20° .

Although this last problem is a troublesome one, it can actually be used as a powerful tool in identifying the beginning of the S motion. In order to do this, one must construct particle-motion diagrams (commencing at a time known to be earlier than the arrival of S) and identify the beginning of S on the basis of the predicted motion. In selecting a time for S, it is best to use the onset of the SH motion, since the surface SH motion is in phase with the incident motion. This is not true for the surface-SV motion at these distances.

The results of S-wave studies which pertain to the problem of upper mantle structure will now be discussed according to geographic regions, in a manner similar to the discussion of the results of P-wave studies.

2.3.1. NORTH AMERICA. Byerly [51], in his study of northern California earthquakes, found that the travel times of S satisfied a linear relation with distance up to about 20° , with an apparent speed of 4.35 to 4.5 km/sec. He noted that the S_n arrival could be observed beyond 10° , although the false S wave disappeared at this distance.

Gutenberg [9] has summarized the results of his studies of the travel times and amplitudes of S waves recorded in southern California. He found (see Figure 19) that the travel-time curve between 4° and 25° consisted of two branches, of which the first was almost a straight line terminated by a shadow zone at 8° . The second branch began at about 18° with relatively large amplitudes and with a time delay compared to the first branch. The apparent speed of S_n was 4.55 km/sec. A low-velocity channel for S waves in the upper mantle would be associated with the shadow zone.

Lehmann [55] has recently reviewed her studies of S-wave travel times for short epicentral distances. In northeastern America she found that the first branch of the travel-time curve (corresponding to the wave which she calls S_d) is practically linear until it breaks off at 14° . She noted that S_d , which has an apparent speed of 4.63 km/sec, is a clear and well-recorded phase in northeastern America. The second branch (corresponding to the waves which she calls S_r) begins at 14° , where it is approximately 15 seconds later than S_d . Beyond 20° the S_r travel-time curve corresponds to the Jeffreys-Bullen curve [11]. Her travel-time curves and the corresponding ray paths for the portion of the path in the mantle are shown in Figure 20. In this figure it should be noted that the upper surface, containing the point E, refers to the Mohorovicic discontinuity. The S-wave velocity is assumed to increase from 4.60 km/sec just below this surface to a value of 4.65 km/sec at a depth of 120 km (assuming a crustal thickness of 35 km). In the low-velocity layer between depths of 120 and 220 km, the velocity is taken to be 4.30 km/sec. At the base of this layer the velocity is assumed to increase discontinuously to 4.70 km/sec, and then to increase uniformly to a depth of 410 km at which point another abrupt increase from 4.95

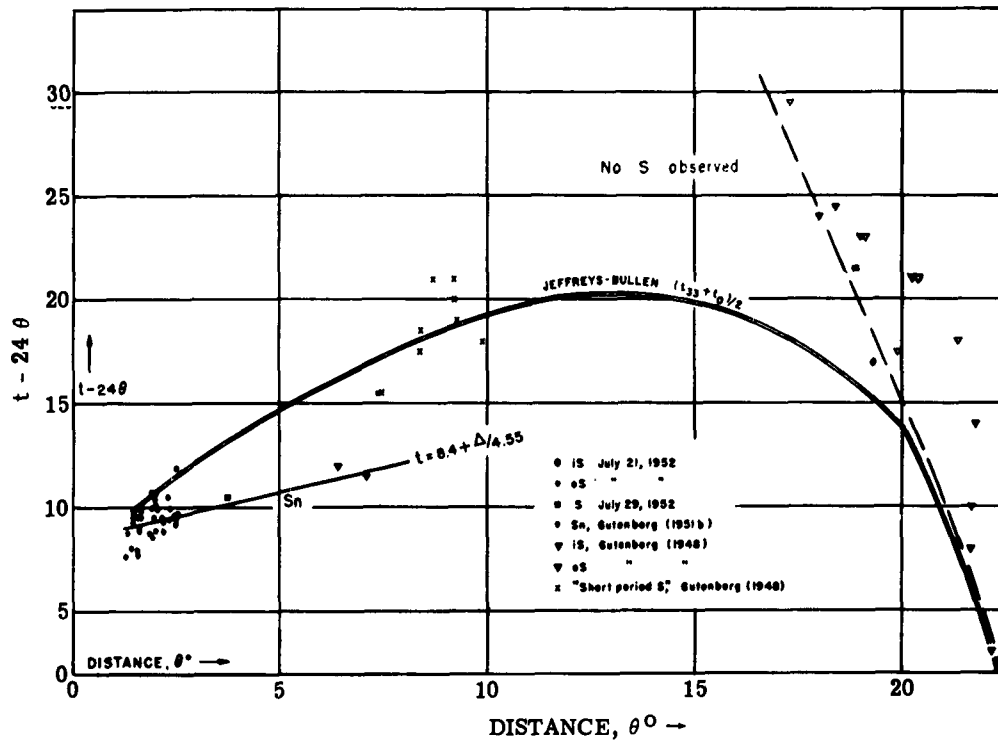


FIGURE 19. REDUCED TRAVEL TIMES OF S FOR SOUTHERN CALIFORNIA [9]

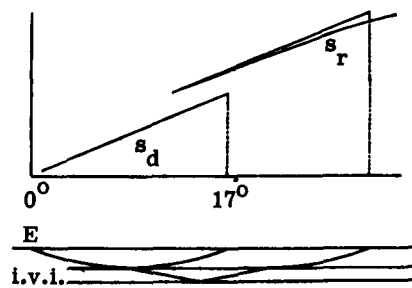


FIGURE 20. RAY PATHS THROUGH THE UPPER MANTLE AND TRAVEL-TIME CURVES WITH THE PRESENCE OF A "LOW-VELOCITY LAYER" [55]

to 5.15 km/sec is assumed to take place. It is possible to make minor adjustments in this model and still satisfy the travel-time and amplitude data.

Jeffreys [18] found the slope of the travel-time curve of S to 10° to be 23.66 ± 0.17 sec/deg, which corresponds to an apparent velocity of 4.696 ± 0.033 km/sec.

2.3.2. EURASIA. On the basis of a study of five shallow-focus earthquakes recorded at 35 European seismograph stations, Ruprechtová [41] determined that the amplitude of the horizontal component of the S motion decreased from 2° to a minimum value at 13° , and then rose again to a maximum value at 22° . This variation of amplitude with distance was similar to that for the P motion, except the minimum and maximum values of P appeared at 11° and 19° , respectively.

Gutenberg [9] reported the results of his study of the S-wave velocities from 12 intermediate focal depth earthquakes in Rumania and Hindu Kush. He found evidence for a low-velocity channel beginning at the top of the mantle, with the minimum velocity of about 4.4 km/sec obtained at a depth of approximately 130 km.

In his study of the travel times of S to distances of 10° , Jeffreys [18] found values of 24.28 ± 0.15 and 24.11 ± 0.10 sec/deg for the slope of the travel-time curves for Europe and central Asia, respectively. The corresponding apparent velocities are 4.576 ± 0.028 and 4.608 ± 0.018 km/sec. He noted that these velocities are much larger than the mean velocity obtained from the Jeffreys-Bullen tables [11] for distances of 2° to 10° , which is 4.429 km/sec.

A comprehensive study of S waves from European earthquakes has been made by Lehmann [55]. She concluded that the S-wave data for shallow European earthquakes can be satisfied by the same upper mantle model as she proposed for northeastern America. She observed, however, that S_d is much smaller amplitude and is less frequently observed in Europe, although on one occasion it was seen at almost 19° .

From a study of the S waves of six Rumanian earthquakes of 130 km focal depth, she found it necessary to change the upper depth of the low-velocity layer from 130 km to 150 km or slightly less. This then required that the value of the low velocity be changed from 4.30 to 4.10 km/sec. The discontinuity in velocity at 220 km, for which she feels there is strong evidence, was retained. She appealed for more special studies of S at small distances and for more adequate instrumentation for recording them.

2.3.3. JAPAN. Gutenberg's [9] velocity variation for S in the upper portion of the mantle beneath Japan (determined from deep-focus earthquakes near Japan, the Marianas, and

the Kurile Islands) is given in the upper portion of Figure 17. The velocity-depth relation shown in the figure would give rise to a low-velocity channel whose upper surface is the top of the mantle. The lower surface of this channel, which is found at that depth for which r/v again attains the value it had at the top of the channel, would be slightly greater than 200 km. These would be the same limits as for the low-velocity channel for P waves in this region.

Nishimura, Kishimoto, and Kamitsuki [45] observed three branches of the S-wave travel-time curve from 10° to 25° . The S_d branch corresponded to the first arrival out to 18° , where it was intersected by the S_{r_1} branch. S_{r_1} was in turn intersected by S_{r_2} at 24° . S_{r_1} and S_{r_2} were first observed at 13° and 19° , respectively. The amplitudes of both S_{r_1} and S_{r_2} decreased with increasing distance. S_d , S_{r_1} and S_{r_2} could be distinguished on the basis of wave period, with S_d having the shortest period (about two seconds) and S_{r_2} the longest. They conclude that there are two surfaces in the upper mantle at which the velocity of S waves increases discontinuously from above to below, although only the first involves a discontinuous change in P-wave velocity.

Fedotov et al. [56] used the observed travel times of both P and S waves originating from earthquakes of focal depth 30 to 120 km to determine the variation of the P- and S-wave velocities with depth in the upper mantle. They found a P-wave velocity of 8.0 km/sec just below the Mohorovicic discontinuity, but average P- and S-wave velocities of 7.8 and 4.5 km/sec for the depth interval of 30 to 100 km. Thus they concluded that the upper mantle contains a low-velocity layer. However, they also found that the ratio of P- to S-wave velocities had a minimum value of 1.71 to 1.72 at a depth of about 70 km, compared to a value of about 1.74 to 1.75 at depths of 30 and 100 km.

It might be noted that a minimum value in this velocity ratio corresponds to a minimum value of Poisson's ratio. Although the authors do not mention the fact, it may be further noted that a minimum value of Poisson's ratio at a depth of 70 km would indicate that the material at that depth has a higher rigidity than at lesser and greater depths, which is just the opposite of that found by most other investigators.

2.3.4. OTHER AREAS. Kogan [57] observed that the arrival times of S waves produced by the United States nuclear explosions in the Marshall Islands were late with respect to the Jeffreys-Bullen tables [11] for a surface focus. She had eleven observations of S waves in the distance interval of 35° to 61° . The time residuals, as computed from the Jeffreys-Bullen tables, varied from 0 to +8 seconds, with seven of the residuals being +4 or +5 seconds. Kogan

concluded that the absence of a marked change in these residuals with epicentral distance suggests that the velocity of S might be somewhat less than that adopted for the upper part of the mantle. She noted, however, that the data were quite limited and that the scatter in the values of the observed residuals might mask a slight dependence of residual on distance, if such a dependence existed. If such were the case, the residuals would not result from incorrect values of the adopted upper mantle velocity, but rather of incorrect values of velocity at much greater depths in the mantle.

Bolt, Doyle, and Sutton [47] reported S_n arrivals out to 11° from the atomic explosions in Australia. They remarked that S_n amplitudes were comparable to those of P_n , but the periods were slightly longer. The S_n times could be fitted by a linear travel-time curve, which gave an apparent speed of 4.75 km/sec.

Cleary and Doyle [48], in their analysis of the east Australian earthquake of May 21, 1961, found that the S_n phase was not clear at most stations. The data from five stations, fitted by a linear curve, gave an apparent velocity of 4.7 ± 0.2 km/sec. They remarked that S was small and appeared to be missing at some stations out to 15° , which suggests the possibility of a layer with low S velocities below eastern Australia, but one which is not as marked as that in southern California.

Jeffreys and Bullen [11], in the preface to their Seismological Table, note that S is not satisfactorily readable in earthquakes of normal focal depth for distances up to 25° , but it is very clear for deep-focus earthquakes. Although S is small for normal-depth earthquakes, they say it can sometimes be recognized by the change in the direction of movement. Their travel times for S for shallow depths of focus and small epicentral distances were derived from observations of S waves produced by deep-focus earthquakes.

3

SURFACE-WAVE STUDIES

3.1 METHODS OF ANALYSIS

The study of long-period surface waves which are influenced by the structure of the earth's mantle is a relatively recent one; most of the significant work has been done during the past ten years. Much important information pertaining to the shear velocity and density of the mantle has resulted from this research.

Surface-wave analysis is concerned with the relationship between group and/or phase velocity and wave period. Explicit expressions can be derived relating these velocities to the period, provided that one specifies the values of the elastic moduli and the density over a suitable range of depth. The usual method of interpreting surface-wave data consists of selecting models of the earth with assigned variation of elastic constants and density (these may be based upon body wave information), deriving a relation between velocity and period for such models, and comparing the results of the computed velocities with the observed data. Another method of interpretation, the inversion method (e.g., Dorman and Ewing [58]), adjusts the values of the parameters of a chosen earth model to minimize the mean square of the residuals of the empirical data with respect to the computed results.

Group velocities are customarily determined by taking the ratio of the epicentral distance to the travel time for a given period wave. Thus they represent an average group velocity over the entire path which they travel.

Several methods are employed to determine phase velocities. Nafe and Brune [59] have summarized these methods, and have included the pertinent equations. It may be noted that the method employing data from a single station gives the average phase velocity between the epicenter and the station in the form of a discrete set of phase-velocity curves. The selection of the correct curve from the set requires a knowledge of the phase velocity at some one particular period, of the initial phase at the source, and of the phase shift introduced by the seismograph. The method employing data from two stations located on a great-circle path passing through the epicenter, or of a single station at which several return waves such as R_2 and R_4 are recorded, yields the average phase velocity along the great-circle path between the stations. Application of this method may also result in a set of phase-velocity curves, from which the correct one must be selected if the observed wave trains have no crests in common. However, if the same wave crests can be correlated between two or more closely-spaced stations, then the phase velocity between the stations is determined uniquely. By using the times of arrival at three stations, one can determine direction of approach as well as phase velocity.

Free oscillations of the earth can also be used to provide phase-velocity data, especially for the very long periods. However, as noted by Brune, Ewing, and Kuo [60], the data from standing waves cannot readily be associated with a given part of the earth.

3.2 RESULTS FROM RAYLEIGH-WAVE STUDIES

Present results of the studies of mantle Rayleigh waves do not lend themselves as readily to a discussion by geographical regions as do the body waves, because the quantity of phase and group velocity data for long-period Rayleigh waves is not great. Since in many cases several

investigators have analyzed the same set of data, it has been decided to review only the most recent paper or papers which discuss a given set of data. Such a decision seems justifiable because improvements and refinements in the methods of interpretation and analysis have been evolving during the past years, and the most recent interpretations appear to be the most reliable ones.

Dorman, Ewing, and Oliver [3] used the group velocities of Rayleigh waves in the period range of approximately 60 to 370 seconds to study the shear velocity distribution in the upper mantle. Their data were taken from Ewing and Press [61 and 62] and Sutton, Ewing, and Major [63]. The first two papers contain group velocities obtained from Pasadena seismograms of earthquakes in India, Japan, Tonga, and Kamchatka, and from Palisades seismograms of the Kamchatka earthquake; these velocities were for Rayleigh waves which had traveled one or more times around the earth. The authors (Dorman et al.) classified these data as "continental" group velocity data, although an inspection of the great circle paths suggests that at least a portion of the paths is oceanic. The data of Sutton, Ewing, and Major are for Rayleigh waves recorded at Suva, Fiji, which had traveled across the North Pacific, and thus pertain to an oceanic region.

Dorman et al. computed phase velocity and group velocity for periods between 50 and 250 to 350 seconds for 11 different earth models, each of which consisted of as many as 50 plane layers. By comparing the observed group velocity data with that computed on the basis of the various earth models, they established conclusively that a low-velocity channel for shear waves is required to explain the data. Both the Lehmann and Gutenberg models (based on S-wave studies) gave an excellent fit to the data of Ewing and Press [61 and 62] up to periods of 200 seconds.

Dorman et al. found that the group velocity data [63] for the deep ocean-basin path across the north Pacific indicated that the upper-mantle structure beneath this region differed significantly from the case previously considered. A satisfactory fit to the observed data out to periods of 200 seconds was provided by modifying the Lehmann model so as to raise the low-velocity layer by about 70 km and to decrease the S-wave velocity by about 0.1 km/sec between depths of 220 and 410 km.

Takeuchi, Press, and Kobayashi [64] and Press and Takeuchi [65] independently observed that the Gutenberg model provided a better fit to the Rayleigh-wave group-velocity data in the period range of approximately 200 to 400 seconds. They noted that the minimum group velocity which is observed at about 230 seconds period (approximately 1200 km wavelength) is mainly due to the sharp increase of shear wave velocity at 400 km depth. Their computed curves were based upon an earth model consisting of plane layers.

Bolt and Dorman [66] provided a thorough discussion of the problem of the dispersion of Rayleigh waves in a spherical, gravitating earth. They computed the phase and group velocities

of the fundamental and first higher-mode waves for periods up to 320 seconds for seven models of the earth.

By comparing the calculations for the spherical, gravitating earth model with those for the flat-earth model of Dorman, Ewing, and Oliver [3], Bolt and Dorman showed that the corresponding group velocities for periods less than 250 seconds differ by less than 1%. Thus the results of the 1960 paper remain valid. However, the phase velocities are significantly higher for computations based on the spherical earth as compared to the flat-earth model. These differences amount to about 2.5% (0.10 km/sec) at a period of 150 seconds and 5% (0.25 km/sec) at a period of 300 seconds.

Figure 21 presents theoretical curves for Rayleigh-wave phase velocities for several earth models. Curves 1 to 5 are for the fundamental mode, and curve 7 for the first higher mode. Curves 1, 2, 3, and 7 are based upon a spherical, gravitating earth model, and 4 and 5 upon a flat-earth model. Combinations of Jeffreys' and Gutenberg's velocity distributions and of Bullen's A and B density distributions are assigned to the various cases, as indicated in Figure 21. The observed phase velocities are plotted as crosses or circles. Crosses refer to data reported in Brune [67] for paths which are approximately 40% continental, and circles to data given in Brune, Nafe, and Alsop [68] for a path which is approximately 60% continental. It can be seen that case 3, based upon the Gutenberg velocity model and the Bullen A density model, gives the best fit to the observed data. Bolt and Dorman note that a modification of case 3, which would have a slightly wider low-velocity channel and slightly lower densities below 600 km, may agree even better with the observations.

Case 3 also provides the best agreement for observed group velocities, as may be seen from Figure 22. In this figure the crosses refer to group-velocity observations of the 1950 Assam earthquake (Ewing and Press [61]) and the 1952 Kamchatka earthquake [62]. The figure also indicates the small difference between group velocities calculated on the basis of flat and spherical gravitating-earth models for the range of periods greater than 100 and less than 270 seconds.

Bolt and Dorman compared the observed free periods of fundamental spheroidal oscillations of the earth excited by the Chilean earthquake of 1960 with those computed by using the Gutenberg velocity and Bullen A density model. The agreement between observed and computed values is better than 0.5% for periods between approximately 240 and 350 seconds. The observed data, taken from the paper of Alsop, Sutton, and Ewing [69], were obtained from pendulum seismographs at Palisades and a strain seismograph at Ogdensburg, New Jersey. Several other groups of investigators also reported periods of the free spheroidal oscillations of the earth produced by the Chilean earthquake. Their values of period agree remarkably well among themselves and with those predicted by the Gutenberg velocity and Bullen A density model. The differences in period

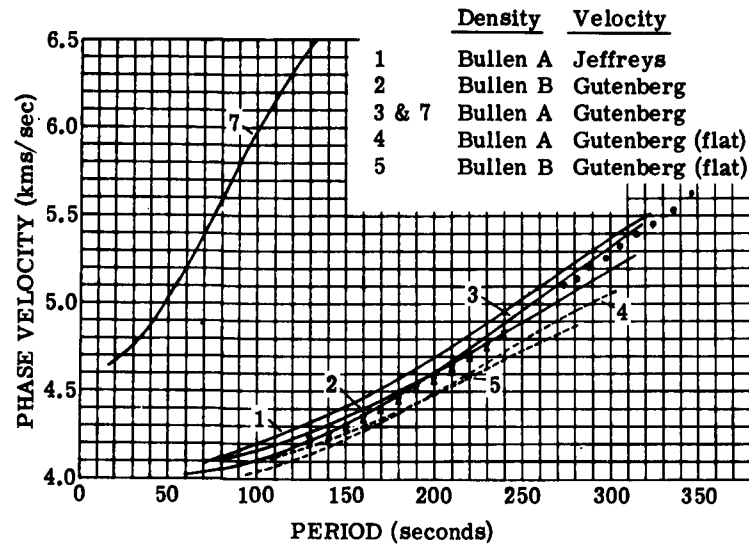


FIGURE 21. THEORETICALLY COMPUTED AND EXPERIMENTALLY OBSERVED PHASE-VELOCITY DATA FOR MANTLE RAYLEIGH WAVES [66]

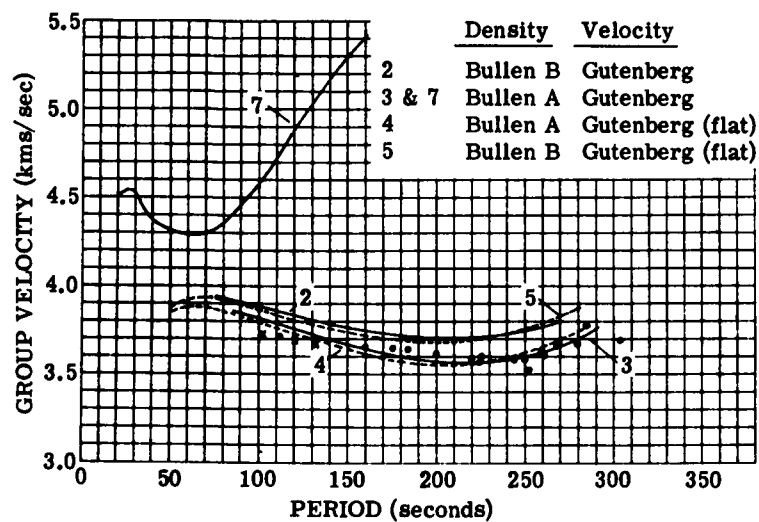


FIGURE 22. THEORETICALLY COMPUTED AND EXPERIMENTALLY OBSERVED GROUP-VELOCITY DATA FOR MANTLE RAYLEIGH WAVES [66]

were in no case as great as 1% for the period range of 240 to 350 seconds, even though they were recorded by a wide variety of instruments. The observations referred to are those of Benioff, Press, and Smith [70] from strain and pendulum seismographs at Pasadena and strain seismographs at Nana, Peru; Ness, Harrison, and Schlichter [71] at Los Angeles from a LaCoste-Romberg tidal gravimeter; and Bogert [71a] at Chester, New Jersey from a pendulum seismograph.

Brune, Benioff, and Ewing [72] extended the period range of observations of group and phase velocity of progressive Rayleigh waves, using seismograms of the 1960 Chile earthquake recorded at Nana, Peru, Isabella, California, and Ogdensburg, New Jersey. Group velocities were observed to periods greater than 1000 seconds, and phase velocities to periods of 650 seconds. Figure 23 contains the observed values of mantle Rayleigh-wave dispersion. In this figure, the solid line and open circles represent data obtained from free spheroidal oscillations, whereas the remaining data were obtained from progressive waves. It is apparent that results based on progressive waves give practically identical values to those obtained from the free oscillations. The maxima in group and phase velocities at periods of approximately 1000 and 650 seconds, respectively, should be noted.

Figure 24, also taken from Brune, Benioff, and Ewing [72], gives observed Rayleigh-wave phase velocities and theoretical curves based upon the Gutenberg velocity and Bullen A density model (solid line) and the Bullen B model (dashed line; see Alterman, Jarosch, and Pekeris [73], for details of this model). The Gutenberg velocity and Bullen A density model gives very good agreement with the observed data.

Ben-Menahem and Toksoz [74] determined Rayleigh-wave phase and group velocities in the period range of approximately 100 to 350 seconds from the Pasadena seismograms of the Mongolian earthquake of 4 December 1957. Phase velocities were computed from the time differences $R_5 - R_3$ and $R_6 - R_4$ as recorded at Pasadena. Their velocities agree very closely with those determined by Brune, Benioff, and Ewing [72].

The preceding discussion of Rayleigh-wave dispersion has not been especially concerned with the paths of the waves. For the most part, the paths were a combination of oceanic and continental, but for no particular ocean and no particular continent. Thus the results may be considered to be valid for an average earth model. It was observed that both phase- and group-velocity data were fairly well satisfied by theoretical calculations based on the Gutenberg velocity and Bullen A density model. One example of group velocity for paths across the north Pacific was presented (data collected by Sutton, Ewing, and Major [63]), and it was noted that the average earth model had to be altered by raising the "low-velocity layer" and also decreasing the S-wave velocity at depths between 220 and 410 km.

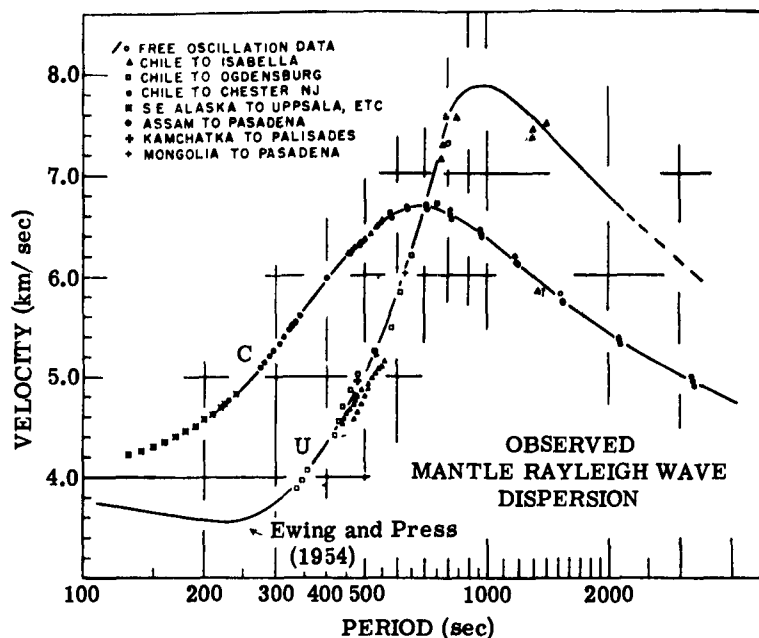


FIGURE 23. PHASE- AND GROUP-VELOCITY DATA FOR MANTLE RAYLEIGH WAVES [72]

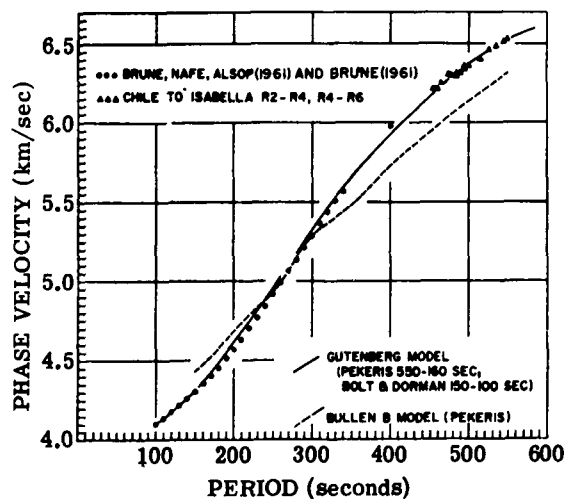


FIGURE 24. RAYLEIGH WAVES PHASE VELOCITIES COMPARED WITH THEORETICAL DISPERSION CURVES. Based on the Gutenberg velocity-Bullen A density model (solid line) and the Bullen B model (dashed line).

Other regional studies of Rayleigh-wave group and phase velocities will now be considered.

Aki [75] studied Rayleigh waves from earthquakes in the north and south Pacific oceans, recorded at Pasadena. He found that the observed seismograms compared satisfactorily with impulse seismograms which were computed using the oceanic model of Dorman, Ewing, and Oliver [3]. Aki and Press [76] noted that the Gutenberg velocity and Bullen A density model did not satisfy the observed Rayleigh-wave data for the Pacific Ocean, but that a reasonable fit could be obtained by simply reducing the velocity in the low-velocity layer from 4.38 to 4.30 km/sec. Thus they furnished an alternative explanation to that of Dorman, Ewing, and Oliver [3] for explaining the dispersion data across the Pacific Ocean.

Kuo, Brune, and Major [77] determined phase and group velocities for Rayleigh waves from earthquakes along the circum-Pacific belt, recorded at Suva, Fiji. Phase velocities between Honolulu and Mt. Tsukuba, Japan, were also determined for two earthquakes originating in Chile and China. Figure 25, showing the great-circle paths, indicates the large percentage of the Pacific Ocean basin that is covered by these paths. Their data (for periods of 20 to 140 seconds) are essentially the same except for the two earthquakes from the southwest Pacific (P 13 and P 14 in Figure 25) whose great-circle paths to Suva cross the Melanesian-New Zealand region.

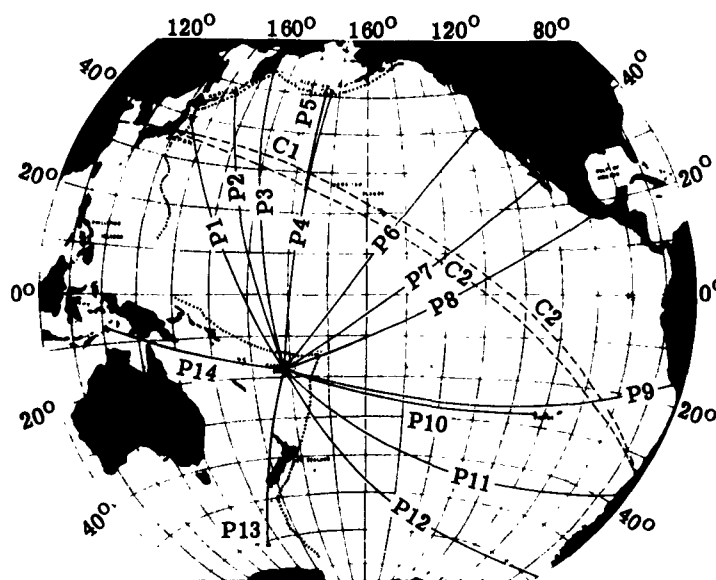


FIGURE 25. MAP SHOWING EPICENTERS AND GREAT-CIRCLE PATHS [77]. Suva, Fiji, is at the point of intersection of most of the great-circle paths.

Both phase and group velocities were significantly lower for these two earthquakes, whose paths lay on the continental side of the andesite line.

In computing phase velocities from the Suva seismograms, Kuo, Brune, and Major assumed a value of the initial phase which made the phase velocity equal to 4.17 km/sec at a period of 120 seconds. They noted that, if it is subsequently found that this velocity is not correct for the Pacific-basin paths, then all their computations will require some small modification. However, their assumption seems reasonable, since Brune [67] found that phase velocities at 120 seconds period from paths that ranged from entirely continental to entirely oceanic (Indian and Atlantic oceans) were all 4.17 ± 0.01 km/sec. They observed that both the modification to the Lehmann model for oceanic paths proposed by Dorman, Ewing, and Oliver [3] and the modification of the Gutenberg velocity and Bullen A density model proposed by Aki and Press [76] to explain Rayleigh-wave velocities across the deep Pacific basin give phase velocities which are greater than their measured velocities in the period range of 50 to greater than 200 seconds. They suggest that the differences could be reconciled either by decreasing the density of the models for depths less than 700 km or by decreasing the shear velocity in the upper mantle.

Aki and Press [76] constructed impulse response seismograms for the Rayleigh-wave portion of waves that had traversed the Indian and Atlantic Ocean basins. They studied the delay in arrival time of the Airy phase by comparing the time difference between the long-period waves and the Airy phase for both the actual-response and the impulse-response seismograms. They assumed that the velocity of the longer-period waves is accurately known, so that the observed time difference corresponded to a delay in arrival of the Airy phase. They showed that this time delay was a linear function of path length in the Indian and Atlantic Oceans, and concluded that the Rayleigh-wave velocities in these oceans are less than for the Pacific basin. The velocities used in constructing their impulse seismograms were based upon the Dorman, Ewing, and Oliver oceanic model [3]. Aki and Press stated that their observations could be satisfied either by assuming lower shear velocities at depths from the Mohorovicic discontinuity to about 50 km (by an amount between 0.1 and 0.2 km/sec) or by a combination of a reduction of shear velocity in this depth interval and a raising of the upper limit of the low-velocity layer.

Brune [67], in his study of the southeast Alaska earthquake of July 10, 1958, noted that the phase velocities over the great-circle path (including the epicenter and the stations Perth and Ottawa) were significantly higher than those for most of the other great-circle paths considered. He concluded that the higher phase velocities probably correlate with the fact that the great-circle path had approximately 270° of its length in deep oceanic areas, primarily the Indian and Atlantic Oceans. He also noted that below 130 seconds period, the phase velocities varied to a considerable extent, depending upon the area traversed. Lower than average phase

velocities in the period range of 80 to 120 seconds were found for those paths which traversed tectonically active regions over a large proportion of their length.

Observations of Rayleigh waves of period greater than the group velocity maximum for strictly continental paths are quite limited. Some results have already been mentioned in the discussion of Brune's study [67] of the 1958 Alaska earthquake. Brune and Dorman [78] presented Rayleigh-wave phase velocities for periods from 2 to 90 seconds for waves which had crossed the Canadian Shield, by correlating phases between Resolute Bay and Ottawa, Halifax, and Pallsades. These data were found by Dorman and Brune [79], using the inversion method, to be satisfied by the following model:

Layer	Thickness (km)	Compressional Velocity (km/sec)	Shear Velocity (km/sec)	Density (gm/cm ³)
1	4.85	5.72	3.40	2.55
2	14.15	5.74	3.59	2.82
3	20.00	6.30	4.04	2.92
4	33.41	8.1	4.77	3.30
5	46.59	8.1	4.95	3.35
6	100.00	8.2	4.43	3.44
7	100.00	8.3	4.69	3.53
8, et seq.,—Gutenberg velocity distributions and Bullen A density distribution				

The most significant point in comparing this model to the Gutenberg velocity and Bullen A density model for the upper mantle is that in the latter the S-wave velocity just below the Mohorovicic discontinuity is 4.65 km/sec; this velocity then decreases continuously to a minimum value of 4.35 km/sec at a depth of 150 km. For the Canadian Shield model, however, the S_n velocity is 4.77 km/sec, and then increases to a value of 4.95 km/sec at a depth of 125 km, where the low-velocity layer first begins.

Caloi [80] observed three types of waves which he called P_a , S_a , and R_a , and identified as propagating through the low-velocity asthenosphere layer. He found that P_a and S_a propagated to great distances with nearly constant velocities of 8.0 and 4.4 km/sec, respectively. Press and Ewing [81] identified similar phases which they called P_n and S_n . The former had speeds of 7.98 to 8.24 km/sec; the latter they divided into two types, having speeds of 4.58 and 4.4 km/sec. They gave a tentative mechanism of "whispering-gallery" propagation in the mantle by means of multiple grazing reflections from the Mohorovicic discontinuity. Khorosheva [82] observed P_a and S_a waves from nine earthquakes, seven of which had focal depths greater than 150 km and four of which had focal depths greater than 400 km. She found velocities of 8.30 and 4.57 km/sec for P_a and S_a , respectively.

Gutenberg [9] accepted Caloi's argument that these waves were guided by the low-velocity channel and concluded that this channel is continuous under both oceans and continents, since

these waves propagated to great distances for all regions of the earth. However, Dorman, Ewing, and Oliver [3] suggested that R_a may be explained as an Airy phase associated with a maximum in the group velocity of the fundamental-mode Rayleigh-wave. They concluded that R_a does not depend upon a low-velocity channel for its existence, although the shape of the dispersion curve in the neighborhood of its period is affected by the shear-wave velocity distribution which determines the channel. Bolt and Dorman [66] noted that S_a could also be explained as an Airy phase, associated with a maximum in the group velocity of the first higher shear mode. Using the Gutenberg velocity and Bullen A density model of a spherical, gravitating earth, they computed that this maximum group velocity would be 4.54 km/sec at a period of 25 seconds. This agrees well with Press and Ewing's description of their Type I S_n wave, which had large vertical motion, periods of 20 to 30 seconds, and an average group velocity of about 4.58 km/sec. No such explanation has yet been forthcoming concerning P_a , but it does not seem unreasonable to speculate that it may be explained as a higher leaking-mode wave.

3.3. RESULTS FROM LOVE-WAVE STUDIES

Much of the present information concerning the observed phase and group velocities of Love waves has been summarized by Brune, Benioff, and Ewing [72]. The results are given in Figure 26. In this figure, the solid lines refer to theoretical calculations for a spherical earth by Pekeris, Alterman, and Jarosch [83] using the Gutenberg velocity-Bullen A density model. The dashed lines refer to similar calculations by Satô, Landisman, and Ewing [84] using the Jeffreys-

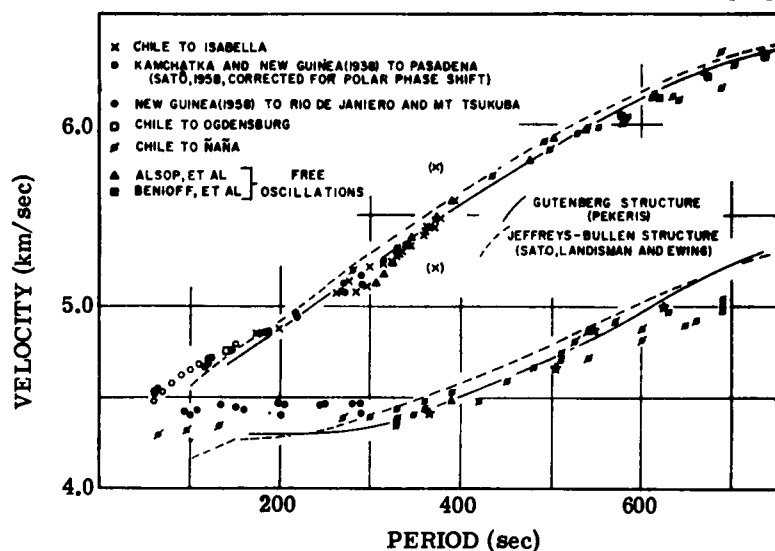


FIGURE 26. THEORETICALLY COMPUTED AND EXPERIMENTALLY OBSERVED PHASE- AND GROUP-VELOCITY DATA FOR LONG-PERIOD LOVE WAVES [72]

Bullen model as presented by Bullard [85]. The latter model employs the Jeffreys S-wave velocity distribution and a density distribution intermediate between the Bullen A and B models. The observed data are those of Satô [86] (which have been corrected for the polar phase shift), Alsop, Sutton, and Ewing [69], Benioff, Press, and Smith [70], and Brune, Benioff, and Ewing [72]. The observed phase velocities (and to a somewhat lesser extent the group velocities) satisfy the Gutenberg velocity and Bullen A density curve for periods greater than about 300 seconds. For periods of less than 200 seconds, the observed phase velocities lie above the theoretical curves. The group velocities for periods less than 300 seconds cannot be obtained directly, but only from a differentiation of the phase-velocity-vs-period curve, since the group velocity is practically constant between 50 and 300 seconds and thus gives a pulse-like surface wave. It can also be observed that the group velocities are significantly higher than the theoretical values for periods less than 300 seconds⁴.

The pulse-like wave resulting from a nearly constant group velocity of 4.4 to 4.5 km/sec is called the G wave. It provides a unique feature on seismograms written by long-period instruments because of its nondispersed shape and its comparatively large amplitude. Press [87] has given a number of determinations of the G-wave velocity; for ocean paths the velocity varied from 4.184 to 4.569 km/sec, and for continental paths from 4.279 to 4.505 km/sec. He noted that the mean velocity for seventeen Pacific Ocean paths was 4.41 km/sec, as it was also for five mixed Asian-North American paths. He concluded that oceanic and continental G-wave velocities probably do not differ by more than 2%.

Satô [86] determined the phase velocity of G waves, using combinations of G_1 - G_3 and G_2 - G_4 waves recorded at Pasadena from the 1938 New Guinea and 1952 Kamchatka earthquakes. The results of his calculations, after correction for the polar phase shift, are included in Figure 26. This paper was significant because in it was presented a method for computing phase velocities of a nondispersed wave by use of the argument of the Fourier transform.

Båth and Lopez Arroyo [88] computed phase and group velocities for Love waves, using the G_1 , G_2 , G_3 , and G_4 waves recorded at Uppsala, Sweden, from the Peru earthquake of January 13, 1960. Phase velocities were determined from the combinations G_1 - G_3 and G_2 - G_4 . The group velocities derived from these phase velocities agreed well with those measured directly, giving

⁴Note added in proof: Recent work by L. Sykes, M. Landisman, and Y. Satô ("Mantle Shear Wave Velocities Determined from Oceanic Love and Rayleigh Wave Dispersion," presented at the October 1962 meeting of the Eastern Section of the Seismological Society of America) shows that there is no disagreement between theoretical and observed phase and group velocities of Love waves. The disagreement mentioned above disappears when account is taken of the earth's sphericity in computing values of the phase and group velocities. There are significant differences in the computed values of these velocities, based upon flat-earth and spherical-earth models, for periods as short as 10 seconds.

velocities of 4.46 km/sec from G_1 - G_3 data and 4.44 km/sec from G_2 - G_4 data for the period range of 40 to 300 seconds.

Brune and Dorman [78] obtained phase velocities for fundamental-mode Love waves in the 4- to 60-second period range for the Canadian Shield area. They also obtained group velocities for these waves and for higher-mode Love waves. These data were incorporated along with Rayleigh-wave and refracted body-wave data by Dorman and Brune [79] in their study of the crustal and mantle structure. The velocities and densities of their model have already been presented in the section on Rayleigh waves. It is important to note that both the Love- and the Rayleigh-wave data for the Canadian Shield are satisfied by this model.

3.4. THE QUESTION OF ANISOTROPY

In all the previous discussion it has been assumed that the earth is an isotropic body, so that the characteristic velocities of the P and S waves at a point are the same for all directions about the point and can be specified by two elastic constants and the density.

Anderson [89] has reviewed and extended the theory of wave propagation in a transversely isotropic medium; i.e., one which possesses an axis of symmetry in the sense that all rays at right angles to this axis are equivalent. For such a medium there are five independent constants relating stress intensity to strain, and for any assigned wave normal there are three characteristic velocities of wave propagation. For the special case of the ray having the direction of the axis of symmetry, the problem degenerates to that of a pure compressional wave and a pure shear wave. If the ray is in the plane perpendicular to the axis of symmetry, there are three characteristic wave velocities, corresponding to a P, an SV, and an SH wave. In this case the SV and SH waves have different velocities of propagation. In the general case of wave propagation in an arbitrary direction, there are still three characteristic velocities of propagation, but the motions associated with them cannot be separated into compressional and shear types.

Anderson showed that the modulus which resembles the rigidity modulus for isotropic materials is not the same for Love-wave motion as for Rayleigh-wave motion in a transversely isotropic material. The result holds true for layered anisotropic media in general and will cause different velocity-density models to be inferred on the basis of separate analyses of Rayleigh-wave and Love-wave data, if the assumption of isotropy is made.

Anderson and Harkrider [90] showed that the discrepancies between long-period Rayleigh- and Love-wave dispersion⁵ can be resolved by assigning a 7% anisotropy to the low-velocity zone of the upper mantle. They noted that the effect of density by itself, assuming an isotropic

⁵See, however, Footnote 4.

body, would be insufficient to remove these discrepancies. They also noted that the S_2 wave (Press-Ewing S_n , Type II) could be explained by their earth model as resulting from a plateau in the higher-mode, Love-wave, group-velocity curve at a period of 20 seconds.

For an isotropic material with a constant Poisson's ratio, the ray paths of the P and S waves will coincide, and one can write expressions for the ratios of the observed P, SV, and SH components at a given point that depend only on the radiation pattern of elastic wave energy at the focus, the azimuth and emergence angle of the ray at the focus, and the effective P- and S-wave velocities at the surface of the earth. Vvedenskaya and Balakina [91] have written these relations in the form

$$\begin{aligned} u_P/u_{SH} &= (V_S/V_P)^2 F_1(l, m, n, Az, e) \\ u_{SV}/u_{SH} &= F_2(l, m, n, Az, e) \\ u_P/u_{SV} &= (V_S/V_P)^2 F_3(l, m, n, Az, e) \end{aligned} \quad (2)$$

where u_P, u_{SH}, u_{SV} refer to the P, SH, and SV components of ground motion, V_P and V_S are the P- and S-wave velocities, l, m, n , are the direction cosines of the principal stress system which describes the focal mechanism, Az is the azimuth from the epicenter to the station, and e is the angle of emergence of the ray at the focus. The functions F_1, F_2 , and F_3 should each equal unity for an isotropic material.

Figure 27 contains the results of the observational studies of Vvedenskaya and Balakina. The straight lines parallel to the epicentral distance axis are the expected values of the ratios for an isotropic earth. It can be seen that both graphs involving the SH component of motion are anomalous at epicentral distances of 18° to 20° , 38° to 42° , 51° to 53° , and about 70° and 80° . The corresponding depths of penetration of rays which emerge at these distances are 250 to 500, 900 to 1000, 1200 to 1300, 1800, and about 2200 km.

Vvedenskaya and Balakina demonstrated that weak absorption in a transversely isotropic medium will result in equal attenuation of P and SV waves in all directions, and in directionally dependent attenuation for SH waves with the maximum in the plane perpendicular to the axis of symmetry (horizontal plane for a layered-earth model). This effect, as well as a difference in the angle of refraction of SH waves compared to P and SV waves, produces the greatest relative attenuation of SH waves for those rays whose vertex or deepest point of penetration lies in the transversely isotropic region. Thus the authors concluded that the depth intervals of 250 to 500, 900 to 1000, 1200 to 1300, 1800, and about 2200 km are regions of the mantle that are anisotropic.

Nuttli [92] studied the horizontal component of the particle motion of the S wave for the specific purpose of determining if there was any evidence of anisotropy. He analyzed the S motion

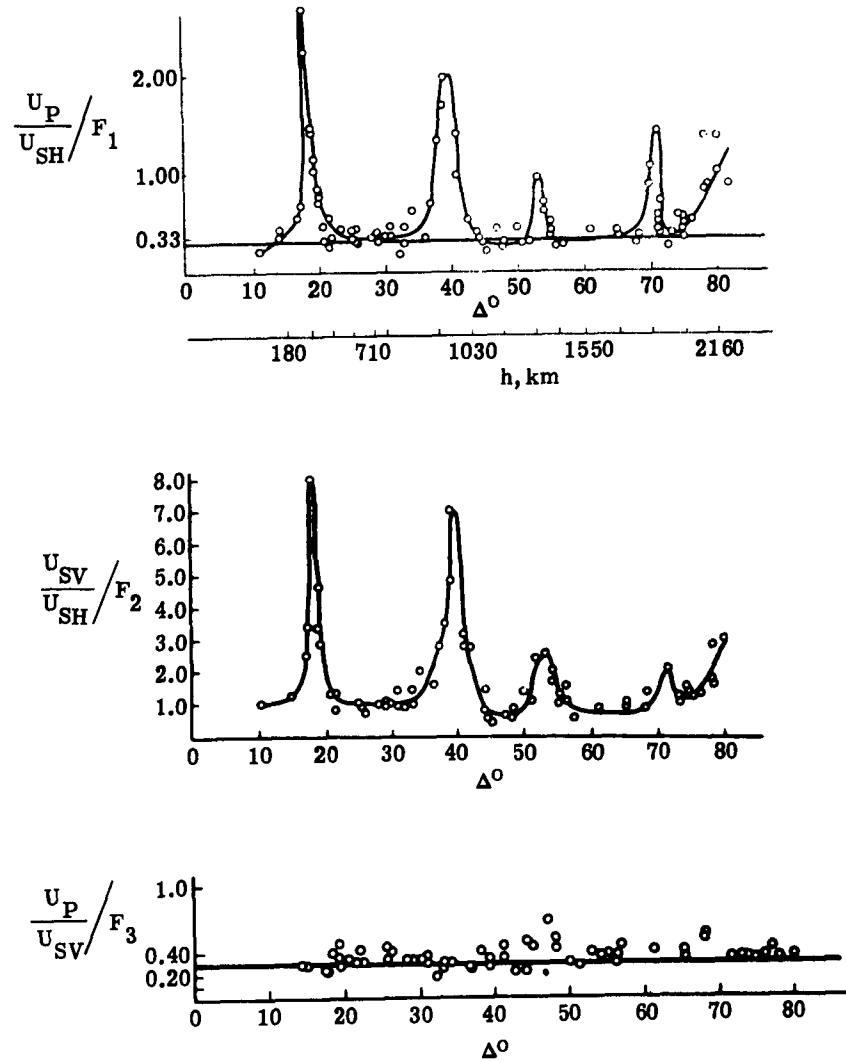


FIGURE 27. VARIATION OF THE RATIOS $(u_P/u_{SH})/F_1$, $(u_{SV}/u_{SH})/F_2$, AND $(u_P/u_{SV})/F_3$ WITH EPICENTRAL DISTANCES [91]

of six earthquakes recorded at Florissant, Missouri, at epicentral distances of 50° to 80° , an interval in which the effect of the earth's free surface does not produce any phase shifts in the SV component of the ground motion. There was no evidence of anisotropy, since the SH and SV components of the waves of these earthquakes arrived simultaneously. In a more recent work, Nuttli and Whitmore [12] reported the results of their analysis of a much larger number of S waves recorded at distances for which the ground motion was expected to be linear. They found three examples of S motion in which the SH component preceded the SV component by one to two seconds. These earthquakes had focal depths of 105, 120, and 125 km. The S waves of the other earthquakes, which were either shallow or at depths of 60, 90, 95, 145, 175, 640, and 655 km, did not exhibit this phenomenon. In these studies both vertical and horizontal component particle motion diagrams were constructed. The horizontal-component diagram, as well as the vertical-SH component diagram, showed the earlier arrival of the SH component. The vertical-SV horizontal component diagram was linear, indicating that the two SV components of the ground were in phase.

4

CONCLUSIONS AND RECOMMENDATIONS

4.1. THE STATE OF THE ART

The two preceding sections were primarily concerned with summarizing the work of the seismologists who have contributed to our knowledge of the earth's upper mantle. Many data have been presented, as well as inferences and conclusions which do not all appear to be mutually compatible. However, some of the areas of disagreement are not so serious as they may first seem, if one limits the conclusions of the individual investigator to the particular geographic region of his work. The difficulties arise when generalizations for the entire earth are made on the basis of data obtained from a limited area or areas. Table I and Figure 28 summarize the information detailed in Sections 2 and 3.

Rayleigh-wave-phase and group-velocity data provide incontrovertible evidence of the existence of a low-velocity channel for shear waves in the upper mantle of the entire earth. The exact velocity-depth variation for shear waves and the extent and position of the low-velocity channel are functions of the regional geology: deep oceanic basin, continental shield, etc. A model of the upper mantle based on the Gutenberg S-wave velocity and Bullen A density distribution appears to provide the best agreement for an average path - i.e., a complete great-circle path which includes both continental and deep oceanic basin portions. These conclusions are also sub-

TABLE I. SUMMARY OF UPPER MANTLE STRUCTURE ACCORDING TO GEOGRAPHICAL REGION*

Region	P Wave Velocity Variation	Based Upon	References	S Wave Velocity Variation	Based Upon	References
North America: Western U.S.	Evidence either for low-velocity layer in upper mantle or for P_n velocities of the order of 7.7 to 8.0 km/sec	Travel times and amplitudes of P waves from earthquakes and nuclear explosions	[6, 16, 18, 25, 32, 33]	Shadow zone for S waves; low-velocity layer in upper mantle, beginning at Moho	Travel times and amplitudes of S waves from earthquakes	[9, 51]
Central and northeastern U.S.	No evidence of shadow zone or low-velocity layer in upper mantle; P_n velocities of the order of 8.2 km/sec and possibly higher	Travel times and amplitudes of P waves from earthquakes and nuclear explosions	[18, 19, 25, 31, 32, 33]	Shadow zone for S waves; low-velocity layer in upper mantle. S wave velocity constant in approx. 100 km of mantle and then begins to decrease	Travel times and amplitudes of S waves from earthquakes	[55]
Canadian Shield				Low-velocity layer in upper mantle; S wave velocity <u>increases</u> to a depth of approx. 120 km and then begins to decrease	Rayleigh and Love wave phase velocities	[79]
Eurasia: Europe and central Asia	No evidence of shadow zone or low-velocity layer in upper mantle; P_n velocities of 8.1 to 8.2 km/sec	Travel times of P waves from earthquakes and Heligoland explosion	[18, 35, 36, 37]	Similar to central and northeastern America	Travel times and amplitudes of S waves from earthquakes	[55]
Arctic-North Atlantic Ocean	Evidence of low-velocity layer of 7.4 km/sec from Moho to 150 km depth underlying the mid-Atlantic ridge and adjacent areas; P_n velocities of 7.7 to 7.8 and 8.0 km/sec in adjacent western and eastern basins of the North Atlantic	Travel times of P waves from earthquakes; P wave refraction studies in ocean	[43, 44]			
Hindu-Kush	Some evidence for a low-velocity layer at a depth of about 100 to 200 km	Travel times and amplitudes of P waves from deep-focus earthquakes	[9, 42]	Similar to western U.S.	Travel times of S waves from deep-focus earthquakes	[9]
Other Areas: Japan	Evidence for either low-velocity layer in upper mantle or for P_n velocities of the order of 7.8 km/sec	Travel times and amplitudes of P waves from earthquakes	[9, 18, 45]	Similar to western U.S.	Travel times and amplitudes of S waves from earthquakes	[9, 45]
South America (Peru and adjacent regions)	Evidence for low-velocity layer in upper mantle	Travel times and amplitudes of P waves from earthquakes	[1]			
Australia	No evidence of shadow zone or low-velocity layer in upper mantle; P_n velocity of 8.1 to 8.2 km/sec	Travel times of P waves from nuclear explosions	[47, 48]	Suggestion of a shadow zone for S, but with upper mantle velocities higher than in western U.S.	Travel times and amplitudes of S waves from nuclear explosions	[48]
North Africa	No evidence of shadow zone between 500 and 1500 km	P waves from nuclear explosion in Sahara	[49]			
Average Earth Model:				Gutenberg S velocity — Bullen A density model (see State-of-Art report, Table II, for values of S velocity and density in upper mantle). In order to explain the observed phase and group velocities of Love waves in the period range of 100 to 300 seconds it may be necessary to assume the presence of some anisotropy in the upper mantle	Phase and group velocities of Rayleigh and Love waves	[3, 66, 72, 74, 90]

* Condensed from "Seismological Evidence Pertaining to the Structure of the Earth's Upper Mantle—a State-of-the-Art Report," by Otto Nuttli. For purposes of tabulation the conclusions have been necessarily simplified; see the State-of-the-Art report for more complete details and qualifications. It should be noted that the results of different investigators do not always agree.

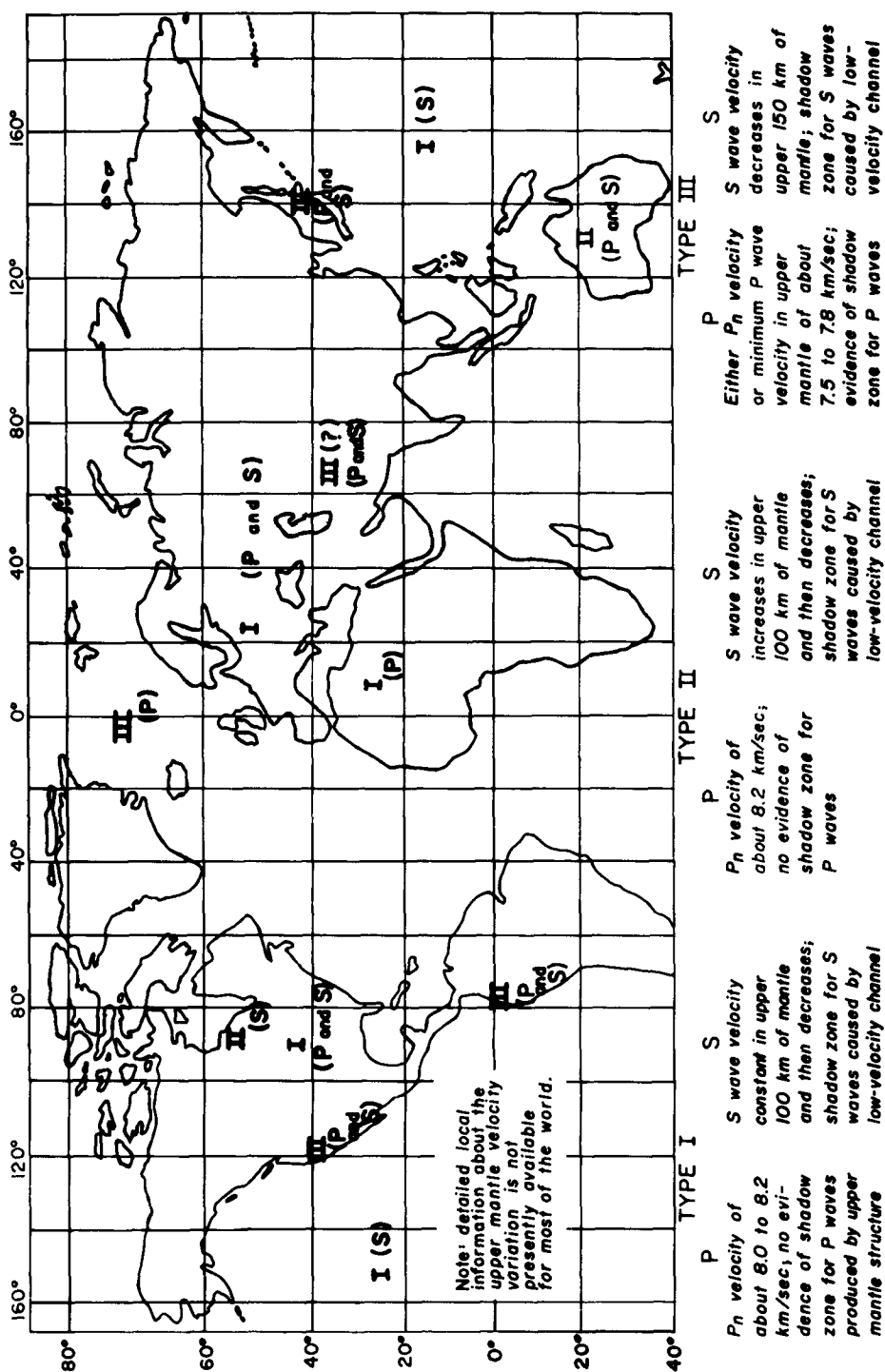


FIGURE 28. GENERALIZED MAP OF P- AND S-WAVE VELOCITY VARIATION IN THE UPPER MANTLE

stantiated by the study of the S-wave travel times and amplitudes, almost all of which establish the presence of a shadow zone corresponding to the existence of a low-velocity channel. In such tectonically active regions as southern California, this channel is more marked than in shield areas, like Canada or Australia. A remarkable agreement exists between Lehmann's model [55] of the upper mantle in northeastern America based on S-wave observations and Dorman and Brune's model [79] for the Canadian Shield based primarily on surface-wave data.

Table II presents the shear wave velocity, density, and rigidity modulus for an average upper mantle, using the Gutenberg S-velocity, Bullen A density model. The data are taken from Alterman, Jarosch, and Pekeris [73]. Values of P-wave velocity and bulk modulus are not included in this table, since the Rayleigh-wave data for all practical purposes are insensitive to these quantities and the S-wave travel times are completely independent of them.

The top of the S-wave low-velocity channel, which occurs at that depth at which $\frac{r}{v} \frac{dv}{dr}$ first equals or exceeds +1, is found to be at the Mohorovicic discontinuity for this model. The bottom of the channel, which occurs at the depth at which the ratio r/v again attains the value it had at the top of the channel, is found to be at a depth of about 260 km.

It may seem curious that the best-fitting model combines the Gutenberg S-wave velocity and Bullen A density models, for the former contains a low-velocity channel whereas the latter was derived from the Jeffreys P- and S-velocity distributions which do not include such a region. This can be explained by the fact that the Rayleigh-wave velocities do not depend in a critical manner on the density distribution, so that minor changes in density would have almost negligible effect. However, it was seen (Bolt and Dorman [66]) that the Bullen A density model gave a better fit to the Rayleigh-wave dispersion data than the Bullen B density model.

Not all the questions connected with the S-wave velocity distribution in the upper mantle have been answered. There remain the important problems of determining the regional variation of the velocity-depth relation, determining whether there are any discontinuous velocity changes in the upper mantle, and determining whether anisotropy provides a unique explanation of the discrepancies between isotropic models derived separately from Rayleigh- and Love-wave data. These topics will be discussed in more detail in Section 4.2.

Seismological evidence of a low-velocity channel for P waves exists for (at most) a few limited areas of the world, namely southern California-Nevada, the Andes of South America, Japan, and the Arctic-North Atlantic Ocean region. On the other hand, the travel times and amplitudes of P do not appear to indicate the presence of any such channel for northeastern America, Europe, central Asia, and Australia. If a region in which the velocity decreases with depth exists in the latter areas, it must begin at a depth well below the Mohorovicic discontinuity and involve only a slight decrease in velocity.

TABLE II. SHEAR-WAVE VELOCITY, DENSITY, AND RIGIDITY MODULUS VALUES IN THE UPPER MANTLE FOR AN AVERAGE EARTH MODEL*

Depth (km)	Shear Velocity (km/sec)	Density (gm/cm ³)	Rigidity Modulus (dynes/cm ²)
0			
	3.55	2.74	3.45×10^{11}
19			
	3.80	3.00	4.33
38			
	4.65	3.32	7.18
50			
	4.62	3.34	7.13
60			
	4.57	3.35	7.00
70			
	4.51	3.36	6.83
80			
	4.46	3.37	6.70
90			
	4.41	3.38	6.57
100			
	4.37	3.39	6.47
125			
	4.35	3.41	6.45
150			
	4.36	3.43	6.52
175			
	4.38	3.46	6.64
200			
	4.42	3.48	6.80
225			
	4.46	3.50	6.96
250			
	4.54	3.53	7.28
300			
	4.68	3.58	7.84
350			
	4.85	3.62	8.52
400			
	5.04	3.69	9.37
450			
	5.21	3.82	10.37
500			
	5.45	4.01	11.91
600			
	5.76	4.21	13.97
700			
	6.03	4.40	16.00
800			

* Based on Gutenberg S-wave velocity-Bullen A-density distribution.

Gutenberg [9] maintained that a low-velocity channel for P waves was world-wide. However, if one considers the data he used (primarily from earthquakes and explosions in California and Nevada and deep-focus earthquakes in South America and Japan), it is not surprising that he arrived at such conclusions. Since all of these regions are tectonically active, they might well be anomalous insofar as the structure of the upper mantle is concerned.

A recent statement of Jeffreys [18] seems pertinent at this point. He noted, "Many seismologists continue to speak of the velocities of P and S as if there were only one for each, valid for all regions. The evidence is now overwhelming that this hypothesis, already in doubt in 1939 [2, p. 504] is false. The lower velocity for P, about 7.8 km/sec, is right for Japan; the higher, about 8.1 km/sec, is right for Europe and Central Asia and Northeast America. Western North America is probably intermediate." Thus there is definite evidence of regional variation in the velocity of the P wave just below the Mohorovicic discontinuity. There are still problems concerning the lower and upper limits of this P_n velocity and of the rapidity of the lateral changes in it. These points will be discussed more fully in Section 4.2.

There is no contradiction in concluding that a particular region contains a low-velocity channel for S but does not give direct evidence of a low-velocity channel for P, in the sense that there is no pronounced shadow zone for P in the distance range of about 700 to 1500 km. This results from the fact that the S-wave velocity depends solely upon the rigidity modulus and the density, whereas the P-wave velocity depends upon the bulk modulus as well. The following two examples were calculated to illustrate this point.

The first example is based upon an upper-mantle model obtained by using the Gutenberg S-wave velocity, Bullen A density (see Table II) and the bulk modulus distribution given by Bullen [4, p. 220]. According to Gutenberg [93, p. 179], there is little disagreement among investigators concerning the value of the bulk modulus in the mantle. The results are given in Table III.

In Table III the quantities V_p and V_s refer to P- and S-wave velocities, respectively, ρ to the density, μ to the rigidity modulus, and k to the bulk modulus. From the data given in the table it can be found that the upper and lower limits of the low-velocity channel for P waves would occur at depths of 60 and 120 km, respectively. The corresponding values for the S-wave channel would be 40 and 260 km. The low-velocity channel is the region in which no rays can bottom or have a vertex. Thus the P-wave travel-time curve would show little evidence of a shadow zone. It would have a slope of about 1/8.0 sec/km out to distances where the waves which had traveled to depths greater than 120 km would arrive. It should be emphasized that the P-wave velocities given in the table are only to be considered in the form of an example, and not as actual values to be found in the earth.

TABLE III. EXAMPLES OF A POSSIBLE P-WAVE VELOCITY DISTRIBUTION IN THE UPPER MANTLE FOR AN AVERAGE EARTH MODEL*

Depth (km)	V_S (km/sec)	ρ (gm/cm ³)	μ (dynes/cm ²)	k (dynes/cm ²)	V_P (km/sec)
40	4.65	3.32	0.72×10^{12}	1.17×10^{12}	8.01
60	4.59	3.34	0.70	1.19	7.99
80	4.49	3.36	0.68	1.21	7.93
100	4.39	3.38	0.65	1.24	7.90
120	4.36	3.40	0.65	1.26	7.90
140	4.35	3.41	0.64	1.29	7.94
160	4.36	3.43	0.65	1.32	7.99
180	4.38	3.45	0.66	1.35	8.04
200	4.40	3.47	0.67	1.38	8.11
220	4.44	3.49	0.69	1.41	8.17
240	4.47	3.50	0.70	1.44	8.23
260	4.51	3.52	0.72	1.47	8.29
280	4.55	3.53	0.73	1.50	8.37
300	4.61	3.55	0.75	1.54	8.47

* Based on Gutenberg S-wave velocity, Bullen A density, and Bullen bulk-modulus distribution.

A second example is given in Table IV, using the S-wave velocity variation found by Dorman and Brune [79] for the Canadian Shield. The density and bulk-modulus values are based upon the same model as those of Table III. For this example, the low-velocity channel for both P and S begins at a depth of 120 km. However, the bottom of the channel for P waves is at 170 km, whereas for S waves it is at 250 km. If one were to fit a straight-line travel-time curve to the first arrivals of P out to about 700 or 800 km (using data resulting from this example), he would very likely obtain an apparent velocity of about 8.2 km/sec. Furthermore, interval velocities based upon time differences between two neighboring stations could get as large as 8.45 km/sec. The latter number is obtained by using the fact that the slope of the travel-time curve is numerically equal to the value of the ratio r/v at the deepest point of the ray path. For this example also there would be little likelihood of being able to observe a shadow zone for P waves.

4.2. PROBLEMS FOR RESEARCH

Seismological problems concerning the earth's upper mantle are ultimately concerned with the determination of the velocity-depth variation for both P and S waves for every region of the earth. Since this is such a vast and formidable problem, it may be desirable to indicate one plan of attack for obtaining the desired information.

TABLE IV. EXAMPLES OF A POSSIBLE P-WAVE VELOCITY DISTRIBUTION IN THE UPPER MANTLE FOR A SHIELD REGION*

Depth (km)	V_S (km/sec)	ρ (gm/cm ³)	μ (dynes/cm ²)	k (dynes/cm ²)	V_P (km/sec)
40	4.77	3.32	0.76×10^{12}	1.17×10^{12}	8.10
60	4.82	3.34	0.78	1.19	8.15
80	4.88	3.36	0.80	1.21	8.24
100	4.93	3.38	0.82	1.24	8.30
120	4.69	3.40	0.75	1.26	8.15
140	4.44	3.41	0.67	1.29	8.01
160	4.46	3.43	0.68	1.32	8.06
180	4.49	3.45	0.70	1.35	8.13
200	4.52	3.47	0.71	1.38	8.19
220	4.54	3.49	0.72	1.41	8.24
240	4.57	3.50	0.73	1.44	8.30
260	4.60	3.52	0.75	1.47	8.36
280	4.63	3.53	0.76	1.50	8.43
300	4.66	3.55	0.77	1.54	8.51
320	4.70	3.57	0.79	1.57	8.57
340	4.74	3.59	0.81	1.60	8.64
360	4.79	3.61	0.83	1.64	8.71
380	4.86	3.62	0.86	1.67	8.81
400	4.95	3.65	0.89	1.71	8.91

* Based on Dorman and Brune S-wave velocity, Bullen A density, and Bullen bulk-modulus distribution.

A study of the regional variation of the phase velocity-period relation of the fundamental and higher-mode Rayleigh and Love waves appears to provide the best hope for obtaining the required information concerning the regional variation of S-wave velocity with depth. Such studies can be made for nonseismic as well as seismic regions, because all that is required is the difference in travel times of long-period waves recorded by seismograph stations in the region of interest. They can also be made in oceanic regions by installing the seismographs on islands.

Since the surface-wave methods are incapable of differentiating between discontinuous and continuous changes in velocity with depth, it would be desirable to study the travel times, and particularly the amplitudes of S waves, for distances out to about 25° for a few selected geographic regions. Although earthquakes will most likely have to provide the source of energy for these waves, it would be desirable to carry out the majority of such studies in regions of relatively low seismicity, because areas of high seismicity will most likely have an anomalous upper mantle structure. The results of the surface-wave phase-velocity and S-wave studies for a given region must be made compatible, of course.

The determination of the regional variation of the velocity of P waves in the upper mantle will have to come primarily from P-wave data, at least with our present state of knowledge. The data required are travel times and amplitudes out to distances exceeding 2000 km. Earthquakes can provide the source of energy for some of these studies, but more complete and useful information can be provided by controlled experiments employing contained underground explosions.

Insofar as the P-wave velocity variation in the United States is concerned, some of these questions could be answered by an experiment involving a contained explosion (either large chemical or nuclear) in the vicinity of northern Louisiana or Mississippi. At least three lines or profiles of temporary seismograph stations could be established, one extending from the source to New Mexico and beyond, the second directly north to Canada and perhaps beyond, and the third in a northeastward direction. The first line would essentially provide a "reverse" profile for the data provided by the GNOME explosion, and thus would furnish information about the effects of any dip or inclination of the Mohorovicic discontinuity between the source and New Mexico. The second would help to decide if the high values of P_n (used in the sense as defined by Herrin and Taggart [33]) as determined for the central and eastern United States from western sources, result from actual high velocity values just below the Mohorovicic discontinuity or if they are produced by rays penetrating rather deeply into the upper mantle in their paths from the Western United States to the seismograph stations. The present writer feels that the latter is a more likely explanation, since it would also simply explain the large amplitudes from GNOME for stations to the east. However, the problem cannot be solved by speculation. The third line would duplicate a portion of those along which GNOME and the Nevada Test Site events were recorded, but in this case for a shorter distance interval. No observable shadow zone for P should be present for any of these three profiles, if the present writer's inferences are correct. As suggested by Lehmann [31], it would be most desirable for such an experiment to have seismographs with a high magnification in the period range of about one to three seconds in addition to the standard Benioffs.

For large portions of the world, including most of the oceanic regions, it does not seem practical at present to obtain a knowledge of the P-wave velocity variation down to depths of 400 or 500 km from direct observations of P waves. Attention should be given to the possibility of inferring the P-wave velocity distribution from the S-wave velocity distribution as obtained from surface-wave analysis. The ability to do this can be tested by independent determinations of P and S velocity for several different geologic regions, which could then be used to determine the density and bulk modulus variation for each region. If the latter quantities are found to be only a function of depth and not of geologic region, then the method used in calculating the examples for Tables III and IV could be employed to determine the P-wave velocity variation. The crucial point concerns knowing whether the decrease of the rigidity modulus with depth in the low-velocity

channel for S in the upper mantle is caused by a change in temperature or a change in chemical composition.

Attention should be given to the possible existence of anisotropic layers in the upper mantle. In particular, studies should be made to determine if transverse isotropy as suggested by Anderson [89] provides the only method of obtaining compatible upper mantle models from Rayleigh- and Love-wave data⁶. More attention should also be given to a study of the S waves from deep-focus earthquake, to delineate those depths for which the SH and SV components do not arrive simultaneously.

4.3. INSTRUMENTATION REQUIREMENTS

Although adequate seismometers and recorders exist to carry out the research suggested in Section 4.2, careful attention should be paid to the selection of the instruments to be certain that the data obtained are of the proper type and of adequate quality.

For example, since the surface waves used in determining phase velocity must have periods up to several hundred seconds, the seismographs must have sufficient long-period response. Since the use of phase-velocity methods on a regional basis involves the time differences between phases over relatively short distances, it is important that the errors in determining the time of arrival of a crest or trough of a wave (or of a Fourier component) be held to an absolute minimum. By way of example, consider two stations 1000 km apart, and a surface wave with a phase velocity of 4.0 km/sec. An error of one second in the time difference will produce an error of 0.02 km/sec in the phase velocity. Figure 22 indicates that such an error is approaching the level of significance when a choice is made between different earth models. If the stations were only 100 km apart, an error of one second would cause an error in the computed phase velocity of about 0.15 km/sec, which suggests that phase-velocity measurements over such small distance intervals are probably impractical when conventional seismograph drum recording speeds are used.

In considering the problem of time control, one must also be concerned with the instrumental phase shift. Although it is relatively simple in theory, it is sometimes difficult in practice to obtain seismographs with identical response characteristics over a wide range of periods. Insofar as the phase shift vs. earth period is concerned, the best results can be obtained by using a seismograph in which both the seismometer and galvanometer are heavily overdamped. However, it is still necessary to calibrate each seismograph directly for the phase shift-period relation if one wishes to determine time differences between a surface wave of period of the order of 100 seconds or greater with an accuracy of at least one second.

⁶See Footnote 4.

In summary, great care must be exercised in the selection of the response characteristics and in the calibration of long-period seismographs if they are to be used for regional phase-velocity studies.

For the purpose of studying the travel times and amplitudes of P and S waves to distances of about 2000 km, it would be very useful to have, in addition to the standard Benioff seismographs, a seismograph with a constant magnification in the period range of about one to three seconds, with the magnification as large as the microseismic background will permit. Not only would such a seismograph provide a better record of the somewhat longer period P and S waves which begin to appear at distances of about 1000 km, but its flat response might also give better amplitude data.

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